THE PRINCIPLES OF AËROGRAPHY

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THE PRINCIPLES OF AËROGRAPHY

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By

ALEXANDER McADIE
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THE PREFACE

Several excellent textbooks on meteorology have been published in this country, the latest having been issued about seven years ago. In the interval since then much new material in connection with the exploration of the upper air has accumulated, which has been published only in scientific journals; and it is thought advisable that an effort be made to present this new knowledge in a convenient form, even if considerably condensed.

Again, the student of to-day is interested in aërography in much the same way that the student of geography is interested in his subject. He is not satisfied with merely locating places, a task essentially mechanical; but goes on to trace the relationship between physiography and the development of communities or nations — truly an educational labor. Thus aërography resembles geography in the larger sense, while meteorology, according to the general acceptation of the term, remains the science of recording diverse atmospheric conditions. The chief purpose of aërography is exploration with a view to utilizing the knowledge gained to insure human safety and to expedite progress.

The present book, therefore, aims to give prominence to recent work that has been done in exploration of the air; such work as that of the first director of the Blue Hill Observatory, the late Professor A. Lawrence Rotch, and his colleague and friend, the late Teisserenc de Bort. Frequent reference is made to the work of Shaw, Dines, Gold, Cave, Hergesell, Assmann, Köppen, Sprung, Süring, Berson, and a host of other modern workers who, in many lands, and often under difficulty, have contributed to this slow conquest of the air.

Another important reason for offering this volume is the desire to further the use of the *c.g.s.* system of units. Throughout the book preference is given to absolute units, in the hope that the student will forget as soon as possible the old, arbitrary, and irrational units. This, it seems to the author, is of importance for at least three reasons: first, the use of these units leads to clear-cut conceptions of the magnitude of the changes, regular or irregular, in pressure, temperature, humidity, and air flow; second, by means of them much time is saved in all computations; and third, they lessen the chance for error in observing and reducing.

More than the usual attention is given to cloud forms and the thermo-dynamics of their formation and dissipation. However, no attempt is made to reproduce *in extenso* weather charts and photographs of common instruments; the former are given in official reports and the latter belong more appropriately to the catalogues issued by instrument makers. Stress is laid rather on modern methods of attack and the practical application of whatever knowledge is already available.

It may be pointed out that the book contains many features not to be found in other textbooks. Some of these are:

1. Results of recent aërological investigations.
2. Introduction of new units.
3. Cloud classification according to origin rather than appearance.
4. Studies of air flow at different levels and the gradient wind.
5. Correlation of abnormal seasons with hyperbars and infrabars.
6. Studies of ice storms, snowfall equivalents, and water supply.
7. Recent floods in the United States.
8. Charts for aviators (Rotch).
9. Variation of ocean currents.
10. Recent knowledge of solar phenomena.

There is also full discussion of thunderstorms and lightning protection; frosts, and the best methods of minimizing the injury therefrom; and other practical problems. The author has tried to speak *with* his readers rather than *at* them, fully realizing that the sum of knowledge is indeed small as compared with that which remains unknown; and he hopes that the book may bring a certain fellowship among those studying the problems of aërography and further the application of knowledge for the good of mankind.

ALEXANDER MCADIE

Blue Hill Observatory
Readville, Mass.
January, 1917

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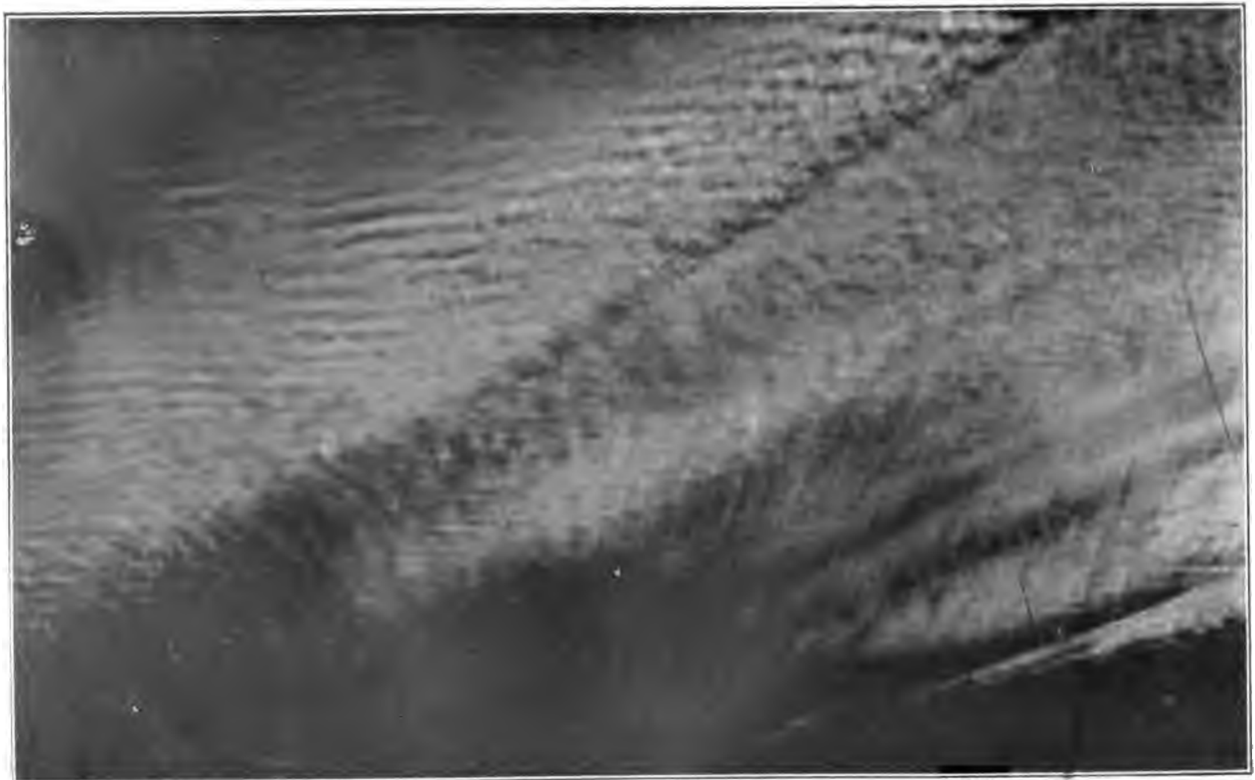


McAdie

CIRRO-CUMULI. HEIGHT APPROXIMATELY 5000 METERS

THE PRINCIPLES OF AËROGRAPHY

CHAPTER I

A Brief History of Meteorology

1. The dawn of meteorology. In a lecture before the Royal Meteorological Society, March 11, 1908, entitled "The Dawn of Meteorology," Hellmann pointed out that, while as a branch of knowledge meteorology is as old as mankind, as a science it is very young.

Ancient weather lore

There is a vast store of weather lore in the Bible, and also in Homer and Hesiod; and there is reason to believe that much of our weather lore dates back to an Indo-Germanic source. It is different from that of Babylonia. There, atmospheric phenomena were associated with the constellations, and a complete system of consequences and combinations was established. This gave rise to astro-meteorology, which became part of the Assyro-Babylonian religion. The Babylonians had a wind rose of eight points and used the names of the four cardinal points to denote the intermediate directions. The Greeks were the first to make regular observations, and as early as the fifth century B.C. *parapegmata*, or peg almanacs, were fixed on public columns. Chiefly these were observations of the wind direction. Anaximander of Ionia (fifth century B.C.) was the first to give a scientific designation to the wind; and it still remains valid. He defined wind as a flowing of the air, ἄνεμον εἶναι ῥύσιν ἀέρος. There is, however, neither a Greek nor a Roman word for wind vane. The Tower of the Winds at Athens probably served more as a public timepiece than as an indicator of weather probabilities.

Soon after these earliest qualitative observations of the weather, came the first that were quantitative. They led to

Early rainfall records

measurements of the rainfall. Such measurements were made in Palestine in the first century A.D. where, as Hellmann points out, "the great influence of rainfall on the crops must have been fully appreciated at an early date and the results of which observations are preserved in the Mishnah, a collection of Jewish religious books apart from the Bible. It seems most interesting that the amount of rainfall then considered as normal for a good crop corresponds pretty closely with that deduced from the modern observations made by Chaplin at Jerusalem; whence it can be inferred that the climate of Palestine has not changed."

Two early physicists

Hellmann further points out that two early physicists, Philo of Byzantium (third century B.C.) and Hero of Alexandria, describe an apparatus which, though primitive, is essentially a thermoscope. Hero's book, on pneumatics, was translated between 1575 and 1592 no less than twice into Latin and three times into Italian. It was studied by Galileo, Porta, and Drebbel, and gave to all three men the idea of constructing a thermoscope and to the last one the impulse of making an experiment on the winds.

The Greeks were also the first to advance theories of meteorological phenomena, and their philosophers had much to say on such matters. A great many speculations were set forth, and by the time of Socrates meteorology was held in low esteem. A new word was coined—μετεωρολέσχης—to designate a mean babbling about sublime things.

Aristotle's treatise on wind

A century later came Aristotle's treatise on wind, the oldest one now in existence. Hellmann says that Aristotle's most distinguished successors, such as Theophrastus and Posidonius, added to it but little. All textbooks of meteorology issued on the continent of Europe before the end of the seventeenth century are exclusively based on Aristotle.

It is of interest to recall that the military expedition of Alexander to the interior of Asia, including India, brought to the Greeks the first knowledge of the monsoon winds; and that the Romans were the first to point out the difference between a continental and a marine climate.

2. The exploration of the air. Exploration and systematic measurement of the various levels of the atmosphere date from the beginning of the twentieth century. Although ancient writers did indeed discuss at considerable length the nature of air, including under the general term *meteorologia* all that was known regarding atmospheric phenomena and much that was essentially astronomical, there was no precise knowledge of the atmosphere as a whole nor any conception of the various motions of the air or its composition. The probable height of the atmosphere may have been apprehended by some of the ancient writers, notably Posidonius. However, very little that is serviceable has come down to us from early times. It is said that in the eleventh century Arabian geometers estimated the probable height of the atmosphere to be 92 kilometers, the computation being made by determining the duration of twilight. Five centuries later this upper limit was redetermined by European astronomers; but not until the middle of the seventeenth century was it known that air could be weighed and that the atmosphere exerted a certain hydrostatic pressure.

Earliest investigations

That such a pressure existed was proved by Evangelista Torricelli, working at Florence in 1643. Torricelli tried the well-known experiment of inverting a tube filled with mercury in a vessel containing the same element, thus demonstrating that a column of mercury *in vacuo* could be balanced against a column of air. He was not concerned about the height of the atmosphere, though this could very easily have been determined, approximately at least, by multiplying the weight per unit volume of mercury (13.596) by the height of the column (76 centimeters at sea level) and dividing by the weight per unit volume of air (0.001293). The result of such a calculation is 7,991 meters, which is the height of the so-called homogeneous atmosphere. The instrument now called the barometer was first known as "Torricelli's tube." The words "barometer" and "baroscope" were introduced by Boyle in 1685.[1]

Torricelli

An experiment of equal significance was that of Blaise

[1] H. C. Bolton, *Science*, April 3, 1903, and A. Lawrence Rotch, *ibid.*, May 1, 1903.

From Tissandier's L'Ocean Aerien

FIG. 1. PASCAL'S EXPERIMENT AT THE TOWER OF ST. JACQUES, 1648

Pascal

Pascal. On September 19, 1648, with the help of his brother-in-law, Périer, Pascal demonstrated that the pressure decreases with elevation. The height of the column of mercury in a vacuum tube was 3 *pouces* (76 millimeters) lower at the summit of the Puy-de-Dôme (a mountain in Auvergne, with an elevation of 1,463 meters) than at the base. Pascal, who had been detained in Paris, soon afterwards showed that a difference could be detected even at an elevation of 50 meters. He noted a difference of two French *lines* (4.5 millimeters) between the readings at the ground and at the upper balcony of the tower of St. Jacques (Fig. 1).

Von Guericke

Boyle

In 1650 Otto von Guericke invented the air pump, proving that air could be weighed and also that the atmosphere exerted pressure. He constructed a huge water barometer, called a *semper vivum*, or *perpetuum mobile*, in which a small floating figure rose and fell with changes in pressure due to atmospheric conditions. It is said that he predicted storms by means of this instrument as early as 1660. In this same year Boyle, in England, published his book entitled *New Experiments Physico-Mechanical Touching the Spring of Air and Its Effects*. Other volumes followed; and in one printed in 1665 there is given a detailed description of changes in the mercurial column and of weather conditions other than mere temperature changes.

Galileo

As early as 1597, Galileo had devised and used at Padua a crude form of thermometer, a modification of which was used in medicine by Sanctorius as early as 1624. The modern form of thermometer did not come into use until the middle of the century.

Meteorology owes much to the little group of nine Florentine investigators, most of them pupils of Galileo and known as the Accademia del Cimento. They made essential improvements in both barometer and thermometer. Descriptions of some of their experiments may be found in the *Saggi di Naturali Esperienze*, published in 1666. The first attempt to establish an international meteorological system of observations was made by Ferdinand II, Grand Duke of Tuscany, in 1654.

In 1650 Southwell, then president of the Royal Society, brought to England the first thermometer, a small one of the Florentine type; and Boyle made a duplicate of it. Within a few years thermometers, barometers, wind vanes, rain gauges, and even dew collectors were in use. Boyle in England, in 1662, and Mariotte in France, in 1676, discovered the law governing the compressibility of air, or of any gas, but the composition of the atmosphere and the true nature of air were not then known.[1]

3. Chemical composition of the atmosphere becomes known. Our knowledge of the chemical composition of the atmosphere dates from the middle of the eighteenth century. Ramsay in his interesting volume, *The Gases of the Atmosphere*, gives in detail the progressive steps by which investigators like Boyle, Mayow, Hales, Black, Rutherford, Priestley, Scheele, Lavoisier, and Cavendish finllay made known what gases existed in the air. Lavoisier, the first successfully to separate oxygen from air, in a pure condition, was also the first to show that water vapor must exist in the atmosphere. He saw clearly that although water is a liquid at ordinary temperatures, it can exist as a vapor and mix with the free gases. Cavendish, that strange and solitary genius, the discoverer of hydrogen, to whom the world of science owes much, published, in 1784, the first of many important papers, entitled *Experiments on Air*. He stated, as the mean of many analyses he had made, the following:

Lavoisier

79.16 per cent of phlogisticated air, or nitrogen;
20.84 per cent of de-phlogisticated air, or oxygen.

These values, says Ramsay, do not differ materially from the best of modern analyses.[2]

[1] Briefly stated, the law governing the compressibility of a gas is this: The volume is inverse to the pressure, provided the temperature remains constant; in other words, the density is in proportion to the pressure. A later and more general form of the law connects volume, pressure, and temperature. It is frequently called the Boyle-Gay-Lussac law, and sometimes, "the equation of state." It is of fundamental importance, as we shall see later.

[2] These give, within small variations,
79.04 per cent nitrogen and argon
20.96 per cent oxygen
after absorption of carbon dioxide, ammonia, and water vapor. (See p. 22 for more detailed statement regarding composition of air.) Carbon dioxide is also

Rayleigh and Ramsay, in 1894, separated from atmospheric nitrogen an inert gas which they called argon. Subsequently other gases, such as helium, neon, krypton, xenon, and niton were discovered. These gases exist in exceedingly small quantities, averaging not more than one part in a million by volume.[1] They are sometimes called the noble gases because of their apparent disinclination to unite with other gases. Niton is an emanation from radium compounds. Coronium is a suspected gas lighter than hydrogen or helium. Its existence is indicated by certain lines in the spectrum of the sun's corona. The name geo-coronium has been used for the extremely rarefied medium at the upper limit of the atmosphere.

The noble gases

4. **Kites and balloons:** *Kites.* Exploration of the free air by means of kites and balloons may be said to have assumed definite proportions at the beginning of the twentieth century. Numerous attempts to use kites and balloons had been made prior to this time, and with a large measure of success; but coöperative, systematic work had not been carried on. Detailed accounts of the various investigations may be found in a report presented to the British Association for the Advancement of Science, at the Winnipeg meeting in 1909; also in an address delivered by Cave before the Royal

found in the atmosphere in small amounts, the weight being about one twentieth of one per cent and the volume somewhat less than this. Hydrogen is found chiefly in the higher levels. At sea level the amount by volume amounts to about one hundredth of one per cent.

[1] The other constituents of the atmosphere are water vapor and traces of nitric acid, also sulphuric acid, bacteria, and dust. While the quantity of water vapor is small and varies with place and time, it is perhaps the most important of all the substances present in our atmosphere and most affects human life. Its functions will be discussed later in the chapters on the formation of the clouds.

Much more importance is now attached to the dust content of the air than formerly. Besides serving as nuclei for condensation, dust plays an important rôle in absorbing and radiating heat. Volcanic eruptions are now known to alter materially the transparency of the air. For example, the violent eruption of Mount Katmai, which occurred on June 6, 7, and 8, 1912, resulted in noticeable dust effects, which have been studied in their relation to transparency by Kimball and others; and with regard to radiation by Abbot, Fowle, and others. The effect of the dust in August was so considerable that the direct radiation of the sun was, by the interposition of the dust cloud, reduced by about 20 per cent at the stations at Bassour, in Algeria, and Mount Wilson, in California. It has been shown that not only the Katmai eruption, but also other great eruptions of former years, have materially decreased the direct radiation of the sun and apparently altered the temperature of the earth. Further discussion will be found in the sections treating of radiation.

Meteorological Society, January 21, 1914. From these, the following abridged history is largely taken.

So far as is known, Wilson of Glasgow was the first to use kites for scientific purposes. In 1749 he raised thermometers, by means of several kites; and on one occasion the top kite reached an amazing height, disappearing into the white summer clouds. Three years later Franklin made his well-known experiments. Kites with thermometers attached were again used in 1821 and in 1836; and in 1847 attempts were made at the Kew Observatory to measure wind velocity and temperature with the aid of kites. In 1865 Nares invented a storm kite for use in carrying life lines ashore in case of shipwreck. Archibald, in 1882, tried to ascertain the increase of wind velocity with elevation, and obtained differential measurements to a height of 300 meters. He also introduced the use of steel wire in place of string, and in 1887 took the first photograph.[1]

First use of kites

Franklin

In 1885 observations of electric potential were made by McAdie at Blue Hill Observatory. The observations were made at moderate elevations by means of kites, the kite string having been wound with fine copper wire. Later, kites were used by Weber in Germany for the same purpose. Eddy, in 1890, devised a tailless kite, a modification of the Malay type, and raised thermometers to moderate heights. He also succeeded in obtaining photographs from cameras thus lifted. But the most important advance in kite work was due to Lawrence Hargrave of Sydney, Australia, who in 1893 or thereabouts invented the box form of kite. Modifications of this form have been used by meteorologists in all countries.

Box kite invented

As a result of the successful use of kites in exploring the upper air, and a report by A. Lawrence Rotch submitted to the International Meteorological Conference at Paris in September, 1896, the Conference recommended that similar investigations be undertaken by the various official weather services of the world. From this time forth the use of the

[1] See also Blue Hill Observatory Reports for 1897; Rotch, *Sounding the Ocean of Air; Quart. Jour. of the Royal Met. Soc.*, April, 1914; and *Nature*, May 28, 1914.

kite spread rapidly. In 1898 daily flights were attempted at eighteen stations in the United States. Notwithstanding the many failures, over a thousand records were obtained, a few of which exceeded 2,000 meters. In 1901 Rotch demonstrated the feasibility of using kites at sea; and in the summer of 1892 Dines flew kites of his own design from steamships. Many investigators have perfected the details of kites and kite meteorographs. Among these may be mentioned Bell, Köppen, de Bort, Berson, Elias, Hergesell, Marvin, Fergusson, and Clayton. Under the direction of de Bort, kites were flown day and night whenever possible for nine months in 1902–1903. The maximum height reached was 5,900 meters. In 1905 de Bort and Rotch organized expeditions for studying, by means of kite records, upper-air conditions over the equatorial Atlantic. The results of these experiments will be given later. The greatest altitude reached by kites—7,044 meters—was attained at Mount Weather, Virginia, on October 3, 1907. During the ascent nine kites were employed.

First use of kites at sea

Greatest altitude reached by kites

Balloons. The first use made of the balloon for meteorological purposes was by Dr. John Jeffries, a native of Boston and graduate of Harvard, living in England. On November 30, 1784, Jeffries made an ascent from London with a French aëronaut named Blanchard, paying the latter one hundred guineas. He carried with him a barometer, a thermometer, a hygrometer, an electrometer, a mariner's compass, and six glass-stoppered bottles filled with distilled water. The bottles were emptied at various levels and sealed; and the samples of air thus obtained were later analyzed by Cavendish. The rate of fall in temperature with elevation as observed by Jeffries (1°C. for every 200 meters) and the decrease of humidity agree fairly well with later determinations. Incidentally, Jeffries, on January 7, 1785, crossed the British Channel in about two hours, starting from Dover and landing in the forest of Guines near the spot celebrated in history as the "field of the cloth of gold."

First use of balloons

Jeffries crosses the British Channel

In 1803–1804 Robertson, a Belgian physicist, made three ascents, from Hamburg and Petrograd, the last for the Russian Academy, with the express purpose of determining the change in the rate of evaporation. A height of 2,430 meters was reached. Biot and Gay-Lussac, on August 24, 1804, under the auspices of the Academy of Sciences, Paris, made an ascent, reaching an altitude of 4,000 meters; and on September 16, Gay-Lussac, at an elevation of 7,400 meters, made certain experiments on magnetism and also collected samples of air. Two noteworthy ascents were made from Paris in 1850 by Barral and Bixio, who demonstrated the great thickness of certain clouds, and showed that in some cases this exceeded 4,000 meters.

Experiments on magnetism

In 1852 the British Association for the Advancement of Science interested itself in aërial exploration, and four ascents were made for the Kew Observatory by John Welsh. The objects were: first, to determine the rate of fall of temperature with elevation; second, to note the change in humidity; third, to make a collection of air samples at various altitudes, and finally to analyze light from the clouds. The thermometers were inclosed in a metal tube through which air was forced. This was the first use of an aspiration thermometer, more perfect forms of which were developed later by Dr. Assmann. Welsh attained heights of from 4 to 7 kilometers, and found that the temperature fell uniformly until at a certain level, varying with the day, the temperature remaining practically constant. Glaisher and Coxwell, from 1862 to 1866, made twenty-eight ascents with the view to determining the rate of temperature fall and the variation in humidity and electrical potential. In cloudy weather the rate of fall was approximately one degree Centigrade for 165 meters, while in clear weather the rate was only half of this. At a height of 8 kilometers no water vapor existed.

Balloons in great altitudes

On September 5, 1862, Glaisher and Coxwell reached a height of 11,200 meters; but as Glaisher was unconscious for a period of about thirteen minutes, and the observations were uncertain, the actual height reached is a matter for doubt. On April 15, 1875, Tissandier, Spinelli, and Sivel, acting for the French Academy, attained an elevation of 8,530 meters; and although

they resorted to oxygen inhalation, the companions of Tissandier were asphyxiated, while he himself was unconscious for some time. On December 4, 1894, Dr. A. Berson attained a height of 9,600 meters, and later (1901) Berson and Suring reached a known height of 10,500 meters (and probably reached 10,800 meters), both being unconscious at the maximum height.

In order to compare the temperatures obtained by Glaisher and others with the values obtained with properly ventilated and protected instruments, simultaneous ascents were made from London and Berlin in 1898. The results showed that the earlier readings of thermometers were in error owing to faulty exposure. Rotch, in ascents made in Paris and Berlin, demonstrated that a self-recording thermometer registered eight degrees higher than a sling thermometer, and the latter two degrees higher than an Assmann aspirated thermometer.

Sounding balloons

In 1893 Hermite and Besançon put into practical form an idea that had been prevalent for some time, namely, the use of small, free balloons capable of lifting light instruments. A varnished paper balloon (*L'Aerophile*) filled with coal gas was first used, but later goldbeater's skin was employed. With such balloons, which weighed about 14 kilograms, a number of ascents were made between 1893 and 1898. A silk balloon (the *Cirrus*, capacity 250 cubic meters, weight 42 kilograms) made eight ascents from Berlin between 1894 and 1895, carrying instruments inclosed in an aspirated tube designed by Assmann. The greatest elevation reached by the *Cirrus* was 18,500 meters, where a temperature of 206° A. was recorded.

International organization

In 1896 the International Meteorological Conference at Paris appointed a committee consisting of De Fonvielle, Hermite, Assmann, Erk, Hergesell, Pomortzeff, and Rotch to organize a series of simultaneous international ascents. The committee also considered the question of uniformity in methods of observation and the interchange of instruments. At the beginning of the twentieth century the work of exploration was officially under way in many lands; also unofficially, as in the case of De Bort and Rotch. Between April, 1898, and 1902, the former sent up no fewer than 258 *ballons-sondes*, which reached heights of

11,000 meters. Similar apparatus was used in the Atlantic expeditions of Rotch and De Bort; also by Hergesell, in 1902. At St. Louis, in 1905, Rotch made the first series of registering balloon ascents in America (Figs. 2, 3, and 4).

At the second meeting of the International Committee, in 1898, it was resolved that (1) thermometers of less thermal

FIG. 2. FIRST ASCENSION IN AMERICA, ST. LOUIS, SEPTEMBER 15, 1904

inertia than those previously used were necessary; (2) efficient ventilation was indispensable; (3) instruments should be tested before the ascents, under circumstances similar to those encountered during the ascents; (4) an aspiration psychrometer suspended at least five feet below the car was the only instrument suitable for manned ascents.

The International Commission for Scientific Aëronautics, under the leadership of Hergesell, has been the main authority in upper-air investigation. The results of simultaneous observations with kites, manned balloons, free balloons, and sounding balloons, and also those made at certain mountain stations, have been published regularly. These international ascents began November 14, 1896, with France, Germany, and Russia participating. Since then, and until the beginning of the European war in 1914, international ascents were made on the first Thursday in each month; on three successive days three times each year; and on six successive days once each year.

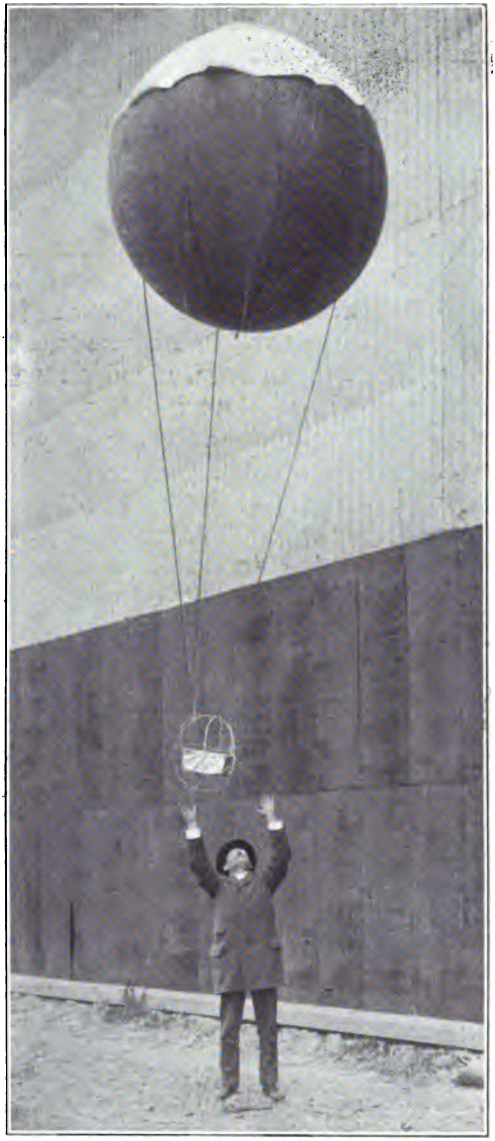


Fig. 3. Launching Ballon-sonde

Congresses have been held as follows: at Strassburg, in 1898; at Paris, in 1900; at Berlin, in 1902; at Petrograd, in 1904; at Milan, in 1906; at Monaco, in 1909; and at Vienna, in 1912. Further meetings were prevented by the European war.

Many observatories in various parts of the world coöperate. The latest report shows the following list: Trappes, Uccle, Soesterberg, Prinsenberg, Pyrton Hill, Limerick, Manchester, Bergen, Christiania, Copenhagen, Pavia, Monticalieri, Milan, Verona, Ferrara, Modena, Florence, Livorno, Vigna di Valle, Monte Cassino, Mileto, Zürich, Friedrichshafen, Stuttgart, Strassburg, Aachen,

International coöperation

FIG. 4. FILLING BALLON-SONDES AT ST. LOUIS

Cologne, Hamburg, Lindenberg, Munich, Vienna, Pola, Trieste, Pavlovsk, Nijni-Oltchedaeff, Sebastopol, Tiflis, Ekaterinburg, Vladivostok, Gizeh, Batavia, Blue Hill, and Mount Weather. Other stations recently established are at Simla under Dr. Walker, at Helwan in Egypt, at Tenerife, and a station in Uruguay. Particularly to be noted is the station in Spitsbergen, where German observers remain not only in

the summer but also through the winter, to study the atmosphere in the Arctic regions; also the station at Batavia, Java, where Dr. Van Bemmelen is doing such excellent work on the winds in the upper air over the equatorial regions (Fig. 5).

"The most complete observatory," says Cave, "for upper-air research is that at Lindenberg. This observatory was founded under the direct personal interest of the Kaiser; and under the direction of Dr. Assmann has carried out an immense amount of work with kites, captive balloons, and registering balloons. Ascents of one sort or another are made every day in the year, and on the international days a large number of ascents are made each day. The Kaiser has also shown his interest in the subject by giving to the International Commission a transportable observatory that, in the first instance, has been erected on the Peak of Tenerife, where the Spanish government now proposes to build a permanent observatory.

The Lindenberg Observatory

"But it is not only in the permanent observatories that work is being done. No expedition for scientific exploration would to-day be complete without some means of studying the upper air. Dr. Simpson worked with balloons in the Antarctic in Captain Scott's expedition; and both Captain Amundsen and the Danish Expedition to Greenland propose to study the upper air.

"Many expeditions have been dispatched for the sole purpose of aërological research. M. Teisserenc de Bort and Professor Rotch chartered a steamer, which, in the years 1905, 1906, and 1907, traversed various parts of the eastern Atlantic, between the temperate zone and the equator, and obtained most interesting results from their observations. The Prince of Monaco made several cruises in his yacht, the *Princess Alice*, in company with Professor Hergesell, notably to the neighborhood of the Canaries and to Spitsbergen.

Aërological research expeditions

"The Lindenberg Observatory organized probably the best equipped expedition. This was sent out for the study of the upper air in tropical Africa. Under the charge of Dr. Berson, twenty-three ascents of registering balloons were made from a steamboat on the Victoria Nyanza from July to September,

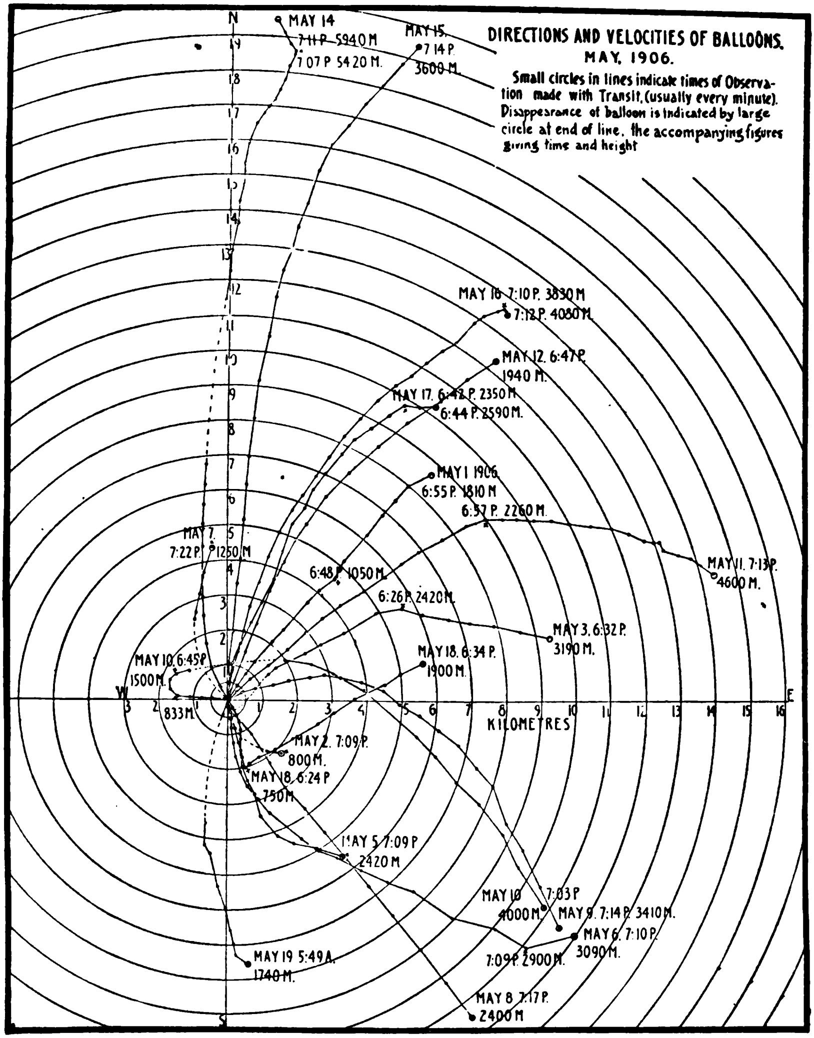


FIG. 5. COURSES TAKEN BY BALLOONS IN MAY, 1906

1908; great heights were reached, and valuable results obtained: much work was also done with kites and pilot balloons. In the international week in July of the same year Professor Palazzo made some ascents with registering balloons from an Italian cruiser in the neighborhood of Zanzibar.

"The most recent aërological expedition is one organized by P. Y. Alexander to study the upper air over the valley of the Amazon; this, too, has been put under the charge of Dr. Berson."

Following the loss of the steamship *Titanic* in April, 1912, the *Scotia* was sent by the British Board of Trade in 1913 as an ice patrol in the North Atlantic and equipped with balloons and kites. This work is now carried on by the U. S. Coast Guard vessels *Seneca* and *Tampa* (formerly the *Miami*).

German Antarctic expedition

Dr. E. Barkow, of the German Antarctic Expedition, obtained during a floedrift journey in Weddell Sea, in the course of a period covering 209 days, many records by kites, captive balloons, and pilot balloons. The greatest height reached was 17,200 meters, February, 1911. Of 256 ascents, 123 were made by kites, 13 by captive balloons, and 120 by pilot balloons. The greatest elevation reached by kites and captive balloons was 2,750 meters; the average, 1,079 meters. The pilot balloons frequently exceeded 10,000 meters, but the average height was 3,598 meters. Jost and Stolberg, 1912–1913, sent up at Godhavn, on the west coast of Greenland, 120 pilot balloons, one of which reached a height of 39 kilometers (24.2 miles).

Records at great altitudes

Unusually high flights in the United States have been those at Huron, S. D., on September 1, 1910, when an elevation of 30,468 meters was reached and a corresponding temperature of 232° A. A minimum temperature of 218° A. was recorded at a height of 15,182 meters. The balloon fell at Castlewood, 105 kilometers east-northeast. On July 30, 1913, at Avalon, Cal., a sounding balloon reached an elevation of 32,643 meters, where the temperature was 231° A. The pressure was as low as 10 kilobars. The lowest temperature, 219° A., occurred at a height of 18,263 meters. Blair,[1] discussing six ascents in which the balloon reached the 20-kilometer level, says that the greatest height reached was 31.6 kilometers on July 9, 1914; and at this height the easterly wind had a speed of 19 meters per second, lowest temperature 211° A. at a height of 16 kilometers.

[1] *Monthly Weather Review*, April, 1916, p. 187.

The lowest temperature thus far recorded in any country was obtained in an ascension at Batavia, Java, December 4, 1913, when, at a height of 17,000 meters, the beginning of the stratosphere in equatorial regions, the temperature fell to 181° A. (−91.9° C.; or −133° F.). In this ascent the balloon reached a height of 26,040 meters; but above 17,000 meters the temperature rose steadily.

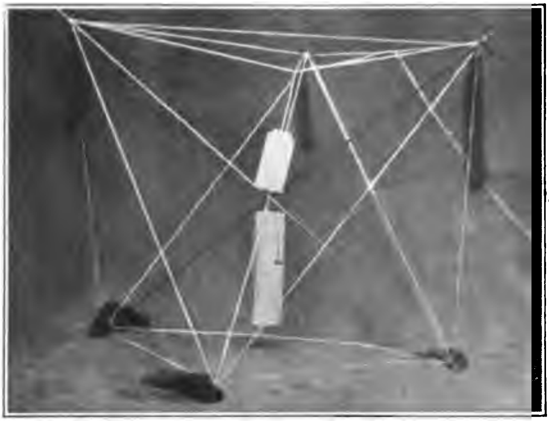


Fig. 6. Dines's Light-weight Meteorograph
During the ascent the meteorograph is suspended inside the bamboo frame or "spider."

Abbot, Aldrich, and Kramer of the Astrophysical Observatory of the Smithsonian Institution have recently made experiments with balloon pyrheliometers with a view to obtaining measurements of solar radiation at great heights.

Balloon pyrheliometers

A radiation record was obtained at about 14,000 meters, and the results indicated that the value of the solar constant (1.93 calories) previously obtained would remain the same. In July, 1914, records were obtained at Fort Omaha at a height of 25 kilometers.

Pilot balloons

Pilot balloons are small free balloons which have been used with great success in studying the wind currents above the surface layers. An account of the methods of observing and the working up of the observations to give the trajectory from which the wind velocity and direction at different heights are obtained is given in a later chapter. (Fig. 6.)

Ten pilot balloons were sent up on different days in 1909 at Blue Hill Observatory, and eleven were sent up in 1910. The pilot-balloon ascension of July 7, 1909, was the first of its kind in the United States. In the most recent Australian soundings self-recording theodolites are employed.

The extreme elevations that have been reached by various means are:

by kites, 7,044 meters, at Mount Weather, Virginia, October 3, 1907;
by manned balloons, 10,500 meters (Berson), July 31, 1901;
by sounding balloons, 37,000 meters, at Pavia, Italy, 1912 (?);
by pilot balloons, height determined by theodolite, 39,000 meters, at Godhavn, 1912–13;
by airplane, 7,950 meters (G. Guidi), November 7, 1916.

The Meteorological Service of Canada sent up 94 balloons from Toronto or Woodstock between January, 1911, and May, 1915. Of these 53 gave records. Most of the balloons traveled in an easterly direction, only three going to the west. The height reached was lower in winter than in summer. The average height of the base of the stratosphere was 11.4 kilometers (7.1 miles) in winter; and 13.4 kilometers (8.3 miles) in summer. The stratosphere had a lower temperature in summer (211° A.) than in winter (214° A.).

5. Determining the height of the atmosphere: *By observations of meteoric phenomena.* Estimates of the height of the atmosphere based upon observations have been made by many astronomers at different places. The paths of shooting stars, meteors, or bolides thus plotted indicate a possible height of from 150 to 200 kilometers. Measurements of the auroral arc and streamers generally

FIG. 7. PILOT BALLOON AS USED AT BLUE HILL

Pilot balloons were sent up on ten days in 1909 at Blue Hill Observatory, and on eleven days in 1910. The pilot-balloon ascension of July 7, 1909, was the first of its kind in the United States.

give much lower elevations and are not reliable, owing to difficulties of identification of the point of measurement.

Shooting stars, meteors, bolides

Perhaps the best results of the measurements of meteors are those given by the great displays of December 24, 1873, and February 18, 1912. The Astronomical Society of Antwerp established an international scientific body (Le Bureau Central Météorique) for the study of meteors, star showers, bright bolides, and every other body of meteoric form. Nagel at Jena and Hoffmeister at Sonneberg maintained simultaneous observations for this purpose during 1913. There is also an American Meteor Society with headquarters at the McCormick Observatory, University of Virginia.

The twilight arch

The twilight arch may serve as a means of determining the upper limit of the atmosphere. It is well known that sunlight illuminates the upper boundary of the atmosphere for some time after the sun sets; and also in the morning for some time before the sun rises. The angle of the twilight arch, as it is called, is found to vary from 15.5° to 18°. We can readily construct a right-angled triangle in which one side will represent the earth's radius, a second side the earth's radius plus the height of the atmosphere, and the included angle one half of the twilight arch, say 9°. The ratio of the longer to the shorter side is the secant of the inclosed angle, or 1.01245. If we take as the earth's radius at 45° a value 6,367,575 meters, we have

$$r+h=r \text{ secant } 9^\circ,$$
$$h=r \text{ (secant } 9^\circ-1),$$
$$h=6{,}367{,}575\ (0.01245)=79{,}276 \text{ meters.}$$

The height of the atmosphere according to this method is, then, about 79 kilometers, or 50 miles. Owing to refraction, or the bending of the rays as they pass through layers of air of different densities, a large correction is necessary, and the value given above is probably too large by as much as 25 per cent.[1]

An approximate method of determining the height of the atmosphere would be by means of cloud measurements,

[1] For tables giving the duration and intensity of astronomical and civil twilight see H. H. Kimball in *Monthly Weather Review*, Nov , 1916.

although the value obtained for the highest cloud would not necessarily be the limiting value for air, but, rather, water vapor. It is of historic interest to note that cloud heights were determined trigonometrically from two stations as early as 1644 by two priests of Bologna, Riccioli and Grimaldi. These were probably lower clouds, and it is unlikely that values as great as 11 kilometers, the cirrus level in temperate latitudes, were obtained. In the tropics, clouds reach somewhat greater elevations, perhaps as high as 17 kilometers.

6. Distribution of gases in the atmosphere. In many of the earlier treatises on meteorology the atmosphere was assumed to be homogeneous. According to this belief there would be a uniform distribution of the gases according to temperature, pressure, and density. It is now clearly proved that such is not the case. On account of well-marked circulations and continuous departure from any state of rest there can be no such uniformity. In other words, the law of Dalton by which each gas would arrange itself independently of the others is not strictly applicable. The gas constant for the air is not a constant. It varies, as we shall see later, owing to the non-adiabatic character of the atmosphere. It is not correct to treat the atmosphere as a dry, perfect gas. A homogeneous atmosphere would extend to a height of nearly 8,000 meters. If the atmosphere consisted of oxygen only, since the density of oxygen is somewhat greater than that of air, a homogeneous oxygen atmosphere would extend upward a less distance, or about 7,200 meters. Nitrogen, being slightly less dense, would reach a height of 8,200 meters. Carbon dioxide, being denser, would not exceed 5,200 meters. Hydrogen would extend to 115,000 meters. A homogeneous water-vapor atmosphere would have its upper limit at a height of 12,847 meters. Homogeneous atmospheres, however, do not exist in nature.

The atmosphere not a perfect gas

Homogeneous atmospheres

Humphreys gives the following summary[1] of our knowledge of the distribution of the gases in the atmosphere:

[1] *Bulletin of the Mount Weather Observatory*, 1909–1910, Vol. II, p. 63.

"The distribution of the atmospheric gases has several times been calculated[1] according to one or another assumption as to the vertical temperature gradient, as to the relative proportions of the several gases in dry atmosphere, and even as to what gases are actually present.

Distribution of gases

"The subject is of general interest and also, in some particulars, of distinct importance to meteorology. And for these reasons it seemed worth while to recalculate this distribution, since the most recent determinations of the factors upon which it depends differ materially from those formerly assumed.

Volume percentages

"Therefore the accompanying table, calculated according to Ferrel's formula for latitude 45°, and graphically represented in Fig. 8, is based on the following assumptions which correspond, we believe, to approximately average conditions:

"(1) That the several gases, in addition to water vapor, present to an appreciable extent in the atmosphere, and their volume percentages in day air at the surface of the earth, are, as Hann[2] gives them:

Nitrogen	78.03	Hydrogen	0.01
Oxygen	20.99	Neon	0.0015
Argon	0.94	Helium	0.00015
Carbon dioxide	0.03		

"(2) That water vapor is present to the extent of 1.2 per cent of the total gases at the surface of the earth, and that it decreases rapidly with increase of elevation, to an imperceptible amount at or below the level of 10 kilometers.

"(3) That the surface temperature is 284°A.

"(4) That the temperature decreases uniformly, at the rate of six degrees per kilometer, from the surface to an elevation of 11 kilometers, where it is 218°A.

"(5) That beyond 11 kilometers above sea level the temperature remains constant at 218°A.

[1] Ferrel, *Recent Advances in Meteorology*, p. 37 (1886); Dewar, *Proceedings Royal Institution*, London, Vol. XVII, p. 223 (1902); Hann, *Lehrbuch der Meteorologie*, p. 8 (1906).

[2] *Lehrbuch der Meteorologie*, p. 5.

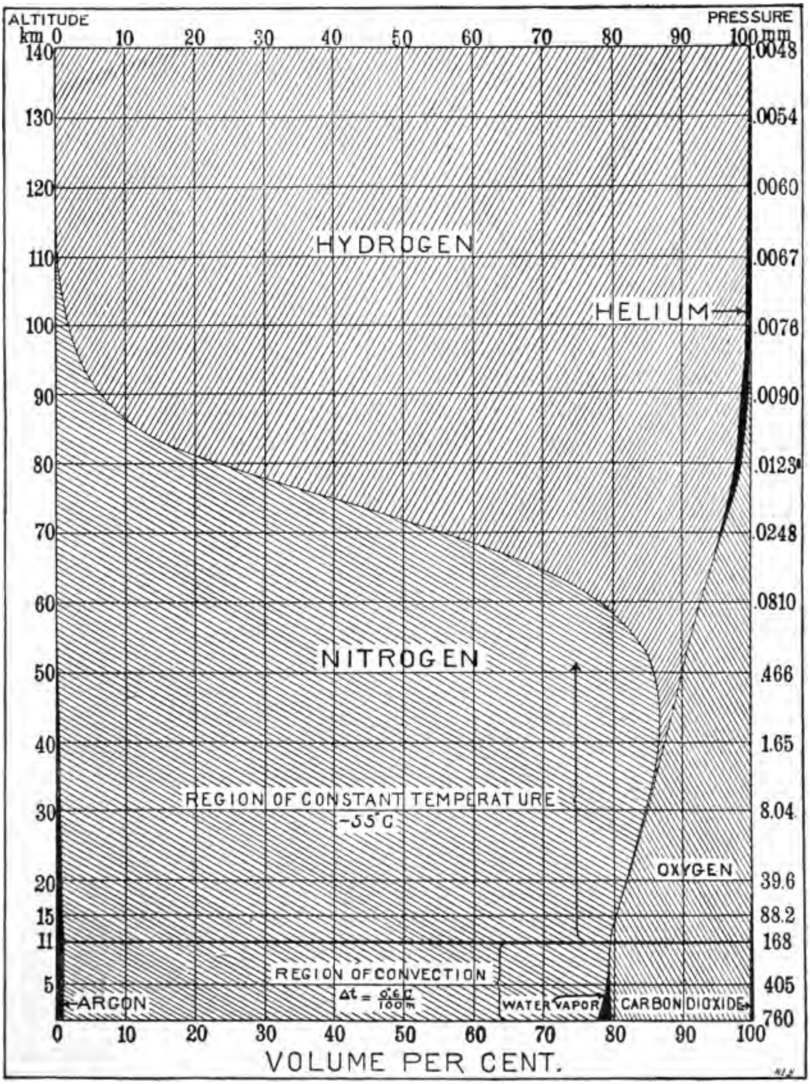


Fig. 8. Distribution of Gases in the Atmosphere — After Humphreys

"(6) That convection, and therefore constant volume percentage of the gases, except as slightly modified by the presence of water vapor, obtains throughout the region of temperature changes; that is, from the surface up to the region of constant temperature.

"(7) That in the region of constant temperature, or that

Convection limits

above 11 kilometers, there is no convection and that in this region the gases distribute themselves according to their molecular weights.

"Fig. 9 must, therefore, be understood to represent both what we know of the lower atmosphere, and what we have reason to believe true of the upper. The one part shows what we are certain of; the other indicates what to look for.

"Probably that fact in regard to the gases of the atmosphere most surprising to the average person is, when its

Small amount of water vapor

immense importance is considered, the relatively small amount of water vapor—an amount which even at the surface of the earth often is no greater than that of argon, and for the total atmosphere scarcely one fourth as great."

7. Molecular weights. In order that we may have some idea of the relative weights of the various gases met within our atmosphere, the following table of molecular weights is given:

Air	28.735
Water vapor	17.880
Oxygen	31.760
Carbon monoxide	27.880
Carbon dioxide	43.760
Hydrogen	2.000
Nitrogen	28.020

The densities are as follows: that of air is 1 at temperature 273°A., and under standard pressure the relative density of hydrogen is 0.0696; of aqueous vapor, 0.6221; of helium, 0.137; of carbon dioxide, 1.529; of argon, 1.3775; of oxygen, 1.1053; of nitrogen, 0.9673. If densities are desired in terms of molecular weight, then

$$\rho = m\,P/KT$$

In this formula m is the molecular weight; P, the pressure of a standard atmosphere or 1,000,000 dynes per square centimeter; K, the radiation energy or approximately 8,000,000 dynes per square centimeter; and T, the temperature in degrees absolute. For air the value is .0013; water vapor, .0008; oxygen, .0014; carbon monoxide, .00125; carbon dioxide, .00197; hydrogen, .00009; and nitrogen, .00126.

CHAPTER II

Units and Symbols

It is not an easy matter to discard terms with which we have long been familiar; and it will be something of a hardship for the present generation to abandon the use of known terms and recognized units, and adopt others in their place. The change to the new notation, however, must be made in order to meet the requirements of the new meteorology now generally called "aërography." The adoption of the new units will result in fewer mistakes, will simplify greatly the compilation of data, and, above all, will lead to definite and precise conceptions of the phenomena of the atmosphere, particularly the various transformations of energy which are manifested in general and local disturbances.

8. The centimeter-gram-second system. This system, commonly called the *c.g.s.* system, is most suitable at the present time and is, moreover, international in character. It is fundamentally the creation of Weber (1852), following Gauss (1832). The first unit, that of length, is the *centimeter*, or hundredth part of a meter. The *meter* is generally defined as the ten-millionth of the meridian passing through Paris. In 1795 the French Republic made this the legal standard of length, and an arc of the meridian extending from Dunkirk to Barcelona was measured by Delambre and Méchain. The standard actually is the distance between two graduations on a platinum-iridium bar at a temperature of 273°A. (0°C.; 32°F.). The bar is preserved in the national archives of France.

Centimeter

The second unit, that of quantity of matter, is the *gram*. It is the thousandth part of the quantity of matter in a standard piece of platinum-iridium called the kilogram prototype. The *kilogram*, or standard of mass, is made as nearly as possible equal to the mass of a cubic decimeter of distilled water at maximum density, 277°A, or 1,014 on the New Absolute scale.

Gram

The third unit is that of time and is the ***mean solar second;*** that is to say, there are 86,400 such seconds in a mean solar day. These units are sometimes called absolute, though, strictly speaking, they are not. Derived from the three units already defined are others: the unit of velocity or the ratio of length to time, the conversion factor being l/t. The unit is one centimeter per second. Another is the unit of momentum, or quantity of motion. This is the mass multiplied by the velocity. The conversion factor is ml/t. The unit of acceleration is one centimeter per second per second, and the conversion factor l/t^2. The unit of area is l^2, the unit of volume l^3, and the unit of density or ratio of mass to volume m/l^3.

Second

The unit most frequently used is that of force, which in the *c.g.s.* system is called the *dyne*, or the force which will impart to the unit mass (a gram) an acceleration of one centimeter per second per second. Force is measured by the rate of change of momentum, and the conversion factor is ml/t^2. Forces are generally measured by weights or the earth's attraction upon given masses. The attraction varies from place to place and with distance from the center of the earth. If we designate by g the acceleration due to gravity, we can say that the weight of a gram at any given place is g dynes. In the old English system, when mass was expressed in pounds, length in feet, and time in seconds, the unit of force was called the *poundal*. In the new system, since a pound is 453.6 grams and a foot is 30.48 centimeters, a poundal would be 13,825 dynes.

The dyne

Work done by a force may result in change of velocity or change of form, the former being a change in kinetic energy and the latter a change in potential energy. The unit of work is the *erg*, and the conversion factor is ml^2/t^2. To raise one kilogram ten meters requires one million ergs. The unit of power is the *watt*, or ten million ergs per second. In the old system the unit was a horse-power, or 550 foot-pounds. If in this system the acceleration of gravity be 32.2 feet per second per second and the foot be 30.48 centimeters, then $17{,}710 \times 30.48 \times 13{,}825$ will give the equivalent value in the new system;

The erg

The watt

and this divided by ten million gives for a horse-power 746.4 watts. A kilowatt therefore is 1,000/746 horse-power (1.34 horse-power).

In 1911 the American Institute of Electrical Engineers adopted 746 watts as the exact value of a horse-power. It is the rate of work expressed by 550 foot-pounds per second at 50° latitude and at sea level. The continental horse-power is 736 watts, or 75 kilogrammeters per second at Berlin, 52° N. latitude.

In the metric system the Greek prefix *deka* indicates ten; *hecto* (which is seldom used), a hundred; *kilo*, a thousand; and *mega*, a million. The Latin prefixes are *deci*, indicating one tenth; *centi*, one hundredth; *milli*, one thousandth; and *micro*, one millionth, in this case a Greek prefix.

Decimalization of angles

The meter was originally intended for use as a geographic unit. Instead of 90 degrees, the earth's quadrant, as previously indicated, was divided into one hundred grades. A grade in turn was divided into a hundred minutes, and a minute into a hundred seconds. The meter is one tenth of the centesimal second, or one ten-millionth of the earth's quadrant. The decimalization of angles has not, however, been generally adopted, though the system has certain advantages. Meanwhile, the arc of 90 degrees on the earth's surface is approximately 10,000 kilometers; and one degree of arc, or 60 nautical miles, is equal to 111.1 kilometers. A nautical mile is equal to 69 statute miles.

Various pressure units

9. The unit of pressure. Unfortunately, in meteorology atmospheric pressure has been represented in three different ways: first, in units of height of a column of mercury *in vacuo*,—as 760 millimeters, or 29.92 inches; second, in units of weight,—as in speaking of the equivalent weight of the atmosphere (1,033.3 grams per square centimeter, or 10,333 kilograms per square meter, or 14.66 pounds per square inch, or 2,111.2 pounds per square foot); and finally, in units of force.

Forces are usually measured by weights, or the earth's attraction upon given masses. This attraction varies from place to place with distance from the center of the earth.

If we designate by g the acceleration due to gravity,[1] then the weight of a gram at any given point is g dynes. The value of g is given by the formula of Bowie,

$$g_o = 978.039\ (1 + 0.005294 \sin^2\phi - 0.000007 \sin^2 2\phi)$$

in which g is the value at sea level and ϕ the latitude.[2] It is now proposed to measure atmospheric pressure in units of force. It has been suggested that the term "Newton" be used as the unit of force 100,000 dynes, and the term "Pascal" for the absolute atmosphere, or 1,000,000 dynes. The unit of force, then,

"Newton"; "Pascal"

NORMAL ACCELERATION OF GRAVITY AT SEA LEVEL IN CENTIMETERS PER SEC. PER SEC.

ϕ = latitude of place

ϕ	0	1	2	3	4	5	6	7	8	9
pole	983.21	....	....	....	....	......	....	....	....	....
80°	983.06	.09	.12	.14	.16	.18	.19	.20	.21	.21
70°	982.61	.66	.72	.77	.82	.87	.91	.95	.99	3.03
60°	981.91	.99	2.07	.14	.22	.29	.35	.42	.49	.55
50°	981.07	.16	.24	.33	.42	.50	.59	.67	.76	.84
40°	980.17	.26	.35	.44	.53	980.62	.71	.80	.89	.98
30°	979.32	.40	.48	.56	.65	.73	.82	.90	.99	0.08
20°	978.64	.69	.76	.82	.89	.95	9.02	.10	.17	.25
10°	978.19	.22	.25	.29	.33	.38	.42	.47	.52	.58
Equator	978.03	.03	.04	.05	.06	.07	.09	.11	.13	.16

is the dyne; and the unit of acceleration, the gal.[3] A dyne is approximately 1/980 of the weight of a gram.

Actual measurements of the intensity of gravity have been

[1] "Gravity" is the term used for the phenomenon of weight, or of the acceleration of a body falling to the earth; and at any place it is the resultant of the earth's attractive force, "gravitation," and the centrifugal force due to the earth's rotation. This distinction between "gravity" and "gravitation" is made by the U. S. Coast Survey.

[2] More generally expressed in a formula of Helmert:

$$g_o = 9.80617\ (1 - .002644 \cos 2\phi + 0.000007 \cos^2 2\phi)$$
$$= 9.8062 \text{ meters at sea level and latitude } 45°.$$

The value 980.665 was adopted in 1888 by the International Committee on Weights and Measures and has since been continued for convenience although it is a conventional standard and not exactly equal to the value at 45°. The best value of the normal acceleration is 980.624.

[3] Whipple has proposed the term "leo" (the last syllable of Galileo) for the unit of acceleration; but—as Klotz points out in *Nature*, August 13, 1914, p. 611—Weichert used the term "gal" (first syllable of Galileo) for this unit in connection with earthquake motion as early as 1909, and it has come into use in seismology. The "milligal" is approximately one one-millionth of g.

made at many points on the earth's surface and these are expressed in accelerations. Normal values, as they are called, are calculated by the general formulæ of geodesy. The intensity of gravity decreases from the pole to the equator; the variation may be readily obtained from the table on p. 28. Barometer readings are frequently reduced to standard gravity by applying the correction for latitude, which is small, and a further correction for elevation.

NEW UNITS

For many years meteorologists felt the need of a unit of pressure that had some definite relation to standard units. Early in 1908 McAdie suggested several modifications of the units and symbols then employed. In the *Monthly Weather Review* for August, 1908, in a plea for new units generally, he described an original method of representing pressure variations in percentages or permillages of a standard pressure. In an extended discussion of the paper in the *Monthly Weather Review* for March, 1909, Köppen suggested that instead of the sea-level pressure, a new base—namely, the pressure represented by the value 1,000,000 dynes—be used. The new pressure base is the pressure at a height of 106 meters (348 feet) above sea level. In other words, instead of using the pressure indicated by 760 millimeters (29.92 inches), which in force units would be 1,013,307 dynes, obtained by multiplying 1,033.291 grams per square centimeter by the normal acceleration of gravity 980.66, we might use the force corresponding to a pressure reading of 750 millimeters (29.53 inches). In April, 1909, Köppen presented to the Aërological Congress at Monaco a strong plea for the use of the new units of pressure.

The new unit of pressure

The first use of the new units in the United States was during 1910, at San Francisco. In Europe, under the stimulation of Shaw, Köppen, and other prominent meteorologists, there has been a ready acceptance of the new units. The use of the centibar and the millibar has become general, and daily and weekly weather reports are published with isobars in millibars and temperatures in the Absolute scale (273° A., 32° F.). On January 1,

The centibar and millibar

1914, the new units were used at Blue Hill Observatory, and on the same date the United States Weather Bureau began the use of these units in a daily map of the northern hemisphere.

Bjerknes in his *Dynamic Meteorology and Hydrography* used the term "bar" as a short and convenient term for the megadyne atmosphere. Unfortunately European meteorologists were not aware that this word had already been defined and accepted in scientific usage, although not very generally adopted. It is of some importance that a short term should be used for the basic unit, and therefore the better usage is to make the bar represent the force of *one* dyne per square centimeter instead of a million dynes. Thus the millibar becomes the kilobar. The following table contrasts the two systems, the New, or American system, and the Old, or European system.

The change from millibar to kilobar

American and European Systems

NEW Chemists and physicists (to be universally used hereafter)	OLD Former aërologists (to be abandoned)	Remarks
1 megabar	1 bar	The absolute atmosphere; equal to 750.1 mm. mercury, or .987 usual sea-level atmosphere. One megadyne per square centimeter acting through one cubic centimeter does one megerg of work.
1 kilobar	1 millibar	One kilodyne per square centimeter.
1 bar	1 microbar	One dyne per square centimeter acting through one cubic centimeter does one erg of work.

For conversion of inches and fractional parts into kilobars see Table 7.

There would be no objection to giving the term megabar or absolute atmosphere some convenient nickname, such as "aer," if megabar is too ponderous. It has been suggested by Professor Richards that for historical reasons the pressure of 10,000,000 dynes (ten absolute atmospheres) might be named after some pioneer in meteorology, as von Guericke or Torricelli, after the analogy of the "watt," "joule," "ampère," etc.; but this need not be insisted on at present.

10. International symbols. The use of certain symbols was agreed upon by the Congress at Vienna in 1873 and these have come into widespread use. These symbols[1] are:

Rain	V Frostwork (rough) forming	T Distant thunder
✱ Snow	Ice coating (smooth) forming	∞ Haze
▲ Hail	Drifting snow	⊕ Solar halo
△ Sleet	← Floating ice crystals	Solar corona
≡ Fog	Gale	Lunar halo
Dew	Thunder storm	Lunar corona
Hoar frost	< Distant lightning	Rainbow
Surrounding country more than half under snow		Aurora

The intensity of a phenomenon is denoted by an exponent, 0 indicating slight, 2, great, and an absence of exponent, moderate intensity.

The time of occurrence is expressed in hours and tenths; morning and afternoon are indicated by A. and P., respectively; midnight and noon by 12 P. and 12 M., respectively, the hours being counted from 0 to 12, commencing at midnight. The continuance of a phenomenon is indicated by a dash (—).

Maximum and minimum values are denoted by heavy-faced type, except for relative humidity, in which case only the minima are so indicated.

The working meteorologist uses, in addition to the symbols and abbreviations given above, certain tables for dividing by 31 and 29; also keeps conveniently near such data as the number of hours in a year, 8,760 (strictly 8,766 and in a leap year 8,784). Certain astronomical and geodetic data are also useful. These are as follows:

General Geodetic Data

1 cm. equals .3937 inch, or .0328 feet. (1 mm. is roughly .04 inch.)
1 cm.^2 equals .155 square inches.
1 cm.^3 equals .061 cubic inches.
1 cm. per second equals .0224 miles per hour, or .0328 feet per second.
The equatorial radius of the earth is 6,378,388 ± 18 meters, or 3,963 miles.
The polar semi-diameter of the earth is 6,356,909 meters, or 3,950 miles.

[1] Professor C. F. Talman in *Monthly Weather Review*, May, 1916, p. 265, discusses in detail the various meteorological symbols used throughout the world.

The reciprocal of flattening is 1/29.4.
The circumference of the equator is 40,076,000 meters, or 24,902 miles.
The perimeter of the meridian ellipse is 40,008,600 meters, or 24,860 miles.
The area of the earth's surface is 196,940,000 square miles, or 510,-044,000 square kilometers.
The area of the ocean is approximately three quarters that of the whole earth's surface.
The mass of the earth is $5{,}984 \times 10^{24}$ kilograms, or 6×10^{21} tons.
The mass of the atmosphere is $5{,}263 \times 10^{15}$ kilograms, or 5.8×10^{15} tons.
The mass of the oceans is about 1.3×10^{24} kilograms, or 1.3×10^{18} tons.
The volume of the atmosphere is approximately $4{,}080 \times 10^{15}$ cubic meters; and since a cubic meter of dry air weighs 1.293 kg. the approximate weight of the atmosphere is $5{,}263 \times 10^{15}$ kgs. This is 1/1,125,000 of the mass of the earth.
The mean density of the earth is 5.52.
The mean density of the surface is 2.67.
The mean density of the ocean is 1.03.
A mean solar day is 24 hours, 3 minutes, 56 seconds sidereal time.
A sidereal day has 86,164 seconds, or 23 hours, 56 minutes, 4 seconds mean solar time.
A sidereal year has 365.26 mean solar days.
The mean distance from earth to sun is 149,500,000 km., or 92,900,000 miles.
Solar parallax, 8.796″; lunar parallax, 3,422.68″.
Sun's diameter, 1,392,000 km., or 865,000 miles.
The mean distance from earth to moon is 384,399 km., or 238,854 miles, or 60.3 terrestrial radii.
The velocity of light is 299,870 km. per second (186,300 miles).
The time required for light to traverse the mean radius of the earth's orbit is 498.8 seconds.

Usage of Certain Letters

ϕ = latitude
λ = longitude
ρ = density
k = ratio of specific heats
γ or g = gravity
ω = angular velocity of the earth
θ = temperature
π = 3.14159265
P and B = barometric pressure
T = temperature on Absolute scale
R = gas constant
V (or v) = volume
V and v used also in some equations for velocity
b = bar
kb = kilobar
m/s = meters per second
e = base of natural logarithms or 2.718281828

No uniformity exists in the usage of letters. Unfortunately, too, letters like g, m, p, and v are used with different meanings. An international standard of notation is much needed

CHAPTER III

Temperature Scales

11. The nature of heat. In his book, *The Constitution of Matter*, Professor Ames says:

"There is no word in our language, I think, which is so much used to conceal ignorance as 'heat,' and no word about which there is so much confusion of ideas as 'temperature.' . . . When we investigate the physical differences between a hot body and a cold one, or when we learn by what physical processes we can make a body hotter, we find that its temperature is determined by the average kinetic energy of translation of the molecules of the body, neglecting any regular systematic motion. (Thus in a tuning fork or vibrating string, these molecular motions are not included in the kinetic energy which determines temperature.) In the case of a gas inclosed in a cylinder, if we push the piston in, there will be an increased kinetic energy of the molecules which will appeal to our senses as a rise in temperature. Similarly, if the gas expands, pushing out the piston, it does work, and the average kinetic energy of the molecules decreases; and we all know that when a gas expands its temperature falls, for example, as shown in the formation of clouds.

Heat a form of energy

"Thus if by any process the average kinetic energy of translation is varied, so is the temperature.

"Lord Kelvin many years ago called attention to the fact that it was possible to define temperature in a way that would be entirely independent of the thermometric substance used in the measuring instrument. This he called the 'absolute' temperature system; and also showed that for all practical purposes this system agreed with the system of using for a thermometer a bulb containing hydrogen or nitrogen and measuring the change in volume, the pressure being kept constant as the temperature changed. He further proved

The "absolute" temperature system

that with matter there is a definite minimum temperature, lower than which it is impossible to reduce a body. This is called 'absolute zero'; and for convenience this is taken as the starting point of the absolute scale."

12. The measurement of heat. Heat may be measured in dynamical units or in thermal units. In the former the dimensions are the same as those of energy, and the conversion factor is ml^2/t^2. In other words, the temperature of a body may be considered to be the average kinetic energy of its molecules and be designated by mass multiplied by velocity squared. When measured in thermal units we determine the amount of heat required to raise unit mass of water one degree. Here the conversion factor is $m\theta$; and the heat unit is the gram-calorie, or the quantity of heat which will raise the temperature of a gram of pure water one degree. Since the specific heat of water varies slightly at different temperatures, the value of the gram-calorie is properly one one-hundredth of the total heat required to raise the temperature of a gram of water from 273°A. to 373°A. Some physicists limit the small calorie, or therm, to the quantity of heat which will raise the temperature of a gram of water from 273°A. to 274°A. This unit is called the small calorie (or therm) because engineers find a larger mass of water more convenient for a unit, and so use a kilogram. The large unit is, then, one thousand times greater than the small calorie.

Gram-calorie

Small calorie, or therm

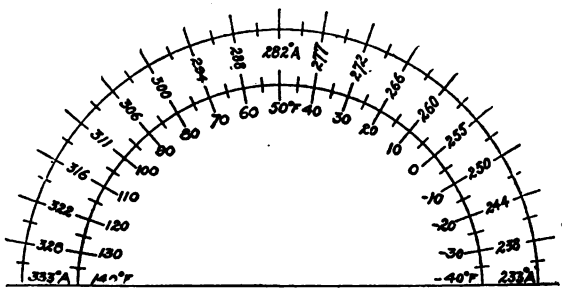
The relative amount of heat required to raise unit mass of any substance one degree compared with water is the specific heat. This varies according as the pressure and volume remain constant. The specific heat of air at constant pressure is 0.24, and at constant volume 0.17. The specific heat of water vapor at constant pressure is 0.47, or nearly twice that of air. The specific heat of water vapor at constant volume is 0.36. In other words, it requires 0.24 gram-calorie to raise the temperature of a gram of air from 273° A. to 274° A., if the pressure remains constant, and twice that amount for a gram of water vapor under constant pressure. The ratio of the

Specific heat of air

specific heat of air at constant pressure to the specific heat at constant volume is 0.2375/0.1683, or 1.41. This value, which is of importance in aërography, is perhaps best represented by the symbol k, although in many treatises the symbol γ has been used. The ratio of the specific heats of water vapor is 0.4734/0.3631, or 1.30, and this can best be designated by the symbol k_1.

Absolute Centigrade scale

13. Temperature scales. It is now desirable to record temperatures on the absolute Centigrade scale, the zero being that of the hydrogen gas thermometer, or practically 273.02° below the zero of the Centigrade scale thermometer, and 459.4° below the zero of the Fahrenheit scale, which, serviceable for many years, has now outlived all usefulness, and is practically discarded in scientific work. It may be recalled that originally this scale



McAdie in "Scientific American"

FIG. 9. CONVENIENT CONVERSION SCALE
This scale gives conversion of temperatures from Fahrenheit to Absolute

ran from −90°, obtained by a mixture of salt and ice, to +90°, the temperature of the human body, the whole making 180°. Later, Fahrenheit made the zero of the scale the lowest temperature of the winter of 1709 at Danzig, 32° the temperature of melting ice, and 212° the temperature of boiling water under constant pressure. The other scales, Réaumur's, the Celsius, and Linnaeus' modification of the Celsius (the modern Centigrade), make the freezing and boiling temperatures of water fiducial points. The Réaumur scale was devised by a French physicist in 1731; and the Celsius by a Swedish astronomer in 1742. He made the boiling point zero and the freezing point 100.

The new zero is that of the hydrogen-gas thermometer. Here the temperature is directly proportionate to the pressure. There is a temperature scale known as the absolute energetic or thermodynamic scale, in which the ratio of any two temperatures is equal to the ratio of the heat absorbed at the one temperature to the heat evolved at the other when the heat is transferred by any reversible cyclical process whatever. Unlike gas or mercury thermometer scales, the definition of temperature does not here involve a relation to any property of any definite substance. Temperatures expressed in this scale are proportional to the pressures given by a constant-volume thermometer filled with a perfect gas. There is a slight difference between the two scales, but no great error in making the zero 273° Centigrade below the temperature of freezing water. The thermodynamic temperature of the ice point is 273.1° A. For rapid conversion of the different scales see Table 9. Or in a general way the illustration Fig. 9 may be used.

Hydrogen gas thermometer

Thermodynamic scale

A modification of the Absolute scale has been proposed by the author; in which starting from the absolute zero, marked 0, the scale divisions have a value of .366 of the Centigrade or Absolute scale division. This makes the reading for the temperature of melting ice 1000. No degree signs are used in the New Absolute scale, as these are to be reserved for angular measurement. The boiling point is 1366.

CHAPTER IV

Thermodynamics of the Atmosphere

14. The specific heat of air. The quantity k, the ratio of the specific heat of air at constant pressure to the specific heat at constant volume, is of much importance in connection with the cooling and heating of the free air. It also enters into the determination of the velocity of sound in air. Newton gave a formula for this velocity, based upon the elasticity of the air; and he considered that the square root of pressure divided by density would give the velocity of propagation of sound. But compression follows rarefaction rapidly in the transmission of sound waves. Laplace pointed out that during such rapid changes of pressure, the pressure did not rise proportionately to the density, as Boyle's law would require. Instead of pressures being inversely proportional to the volumes, that is, $p/p_1 = v_1/v$, they must be inversely proportional to the volumes raised to a certain power, which was k; or $p/p_1 = (v_1/v)^k$. Since $v\rho = v_1\rho_1$, the velocity is equal to the square root of kp/d. The ratio p/ρ (pressure to density) is proportional to the absolute temperature, and therefore the velocity is directly proportional to the square root of the absolute temperature.

The velocity of propagation of sound

At 273° A. the velocity of sound in air is 33,176 centimeters per second. The value of $\frac{k-1}{k}$ is 0.286, and this is equal to the temperature gradient in air multiplied by the gas constant R. The adiabatic temperature gradient for dry air is 9.847° per thousand meters. The value is the same for moist air, provided no condensation occurs; otherwise the heat of condensation partly compensates for adiabatic cooling. The average value for saturated air in the lower strata is not far from 6 degrees per thousand meters, or approximately half the adiabatic rate. Dines, in discussing the vertical distribution of temperature, has shown that under actual conditions the adiabatic gradient holds only for small changes of height and

Adiabatic temperature gradient

not for large changes, and therefore pressure in the higher strata is greater than that given by the formula.[1] The differences for different heights under average conditions are:

1 km.	2 km.	3 km.	4 km.	5 km.	6 km.	7 km.	8 km.
.3°A.	.6	1.0	1.8	2.6	3.5	4.5	6.0

9 km.	10 km.	11 km.	12 km.	13 km.	14 km.	15 km.
7.6	9.7	11.3	13.9	16.7	19.8	25.0

Thus, if a balloon containing air could rise with sufficient rapidity to 15 kilometers, allowing the gas inside to expand adiabatically, the fall of temperature would be only 125°, instead of the 150° given by the commonly used formula.

Mechanical equivalent of heat

The difference between the specific heat of air at constant pressure and constant volume is 0.0692; and this is generally represented by the symbols $C_p - C_v$.[2] This is equal to the heat equivalent of work multiplied by the gas constant. The mechanical equivalent of heat (sometimes called the work equivalent of heat) (J) is 41,840,000, or one calorie equals 4.184×10^7 ergs, or 4.184 joules when not done against gravity. This would raise a gram of air against gravity 426.8 meters.

Emissivity

Joule's equivalent, as it is often called, is connected with the quantity of heat by the equation ML^2T^{-2}, which is equal to JH or $JM\theta$. The conversion factor is $l^2t^{-2}\theta^{-1}$. When heat is measured in dynamical units, J is a number. Emissivity is the quantity of heat given off by a substance per unit time per unit of surface per unit difference of temperature between the surface and the surrounding medium. The conversion factor is $ml^{-2}t^{-1}$. In thermometric units, by substituting l^3 for m the factor becomes lt^{-1}, and in dynamical units, $mt^{-3}\theta^{-1}$.

Latent heat

Latent heat is the ratio of the number representing the quantity of heat required to change the state of a substance to the number representing the quantity of matter in the substance. The conversion factor

[1] Dines, *Quart. Jour. of the Royal Met. Soc.*, July, 1913, p. 187.

[2] See section 16 (p. 41), on dynamical heating and cooling of the air.

is simply the ratio of the temperature units, or θ. In dynamical units the factor is l^2t^{-2}. When a solid is at its melting point, or a liquid at its boiling point, no change of temperature is noticeable when heat is added. There is, however, a change in the internal energy. Conversely, when substances return to their initial condition, as when water vapor condenses, an equivalent quantity of heat is set free or does work of another kind. The heat absorbed or set free in such changes of state, occurring with no apparent change in temperature, is called the "latent heat." Thus we have the latent heat of fusion and of vaporization (or sublimation). The latent heat of fusion of ice, as recently determined, is 79.63 calories. To melt a gram of ice, then, would require theoretically 80 calories. It should be remembered, however, that a gram of ice is by weight a little more than a cubic centimeter; and if pure ice is used only 73 calories are needed. To melt ice containing a percentage of solids such as ice from sea water, a still smaller number of calories will suffice. The latent heat of vaporization is 536 calories; that is, it requires that much heat to change a gram of water into vapor at the boiling point (373°A.).

Latent heat of fusion

Latent heat of vaporization

15. The equation of elasticity. According to the law of Boyle and Mariotte, $pv = p_0v_0$, in which v represents the volume of unit mass of a gas at a pressure p and temperature 273°A. In other words, if the temperature is constant and the pressure be doubled, the volume will decrease one half; and conversely, if the volume be increased twofold, the pressure decreases one half. The law is not true for great or small pressures, but it does hold for the atmosphere when the pressure is about 870 kilobars. At very low pressures the air seems to lose its elasticity.

Volume inversely as the pressure

The law of Charles and Gay-Lussac introduces the temperature as a factor in controlling volume and pressure. It rests upon the experimental fact that the volume of a gas increases 1/273 of its volume at 273°A. for each degree increase of temperature; and it decreases at the same rate. Thus the

Temperature effect on volume and pressure

laws given above may be combined into the following formula:

$$pv = p_o v_o (1 + 1/273\theta),$$

which is another way of saying that at a temperature of 0°A., according to the kinetic theory of gases, there would be no velocity of molecular motion. If, then, we let θ represent temperature on the absolute scale we may write the equation $pv = R\theta$, which means that if the volume remains constant the pressure is proportional to the absolute temperature, and conversely, the volume is proportional to the absolute temperature if the pressure remains constant. As the elastic force or pressure is determined by the mean square of the velocities of the molecules, it follows that the mean is as the absolute temperature.

VALUE OF R, GAS CONSTANT, IN CHARACTERISTIC EQUATION FOR AIR

°A.	Saturation pressure in kilobars	R
250	0.79	2871
260	2.34	2872
270	4.90	2875
273	6.11	2876
275	7.05	2877
280	9.99	2880
285	13.96	2884
290	19.30	2890
295	26.16	2895
300	35.41	2905
305	47.17	2915
310	62.27	2930

The characteristic equation

The characteristic equation is, then, $pv = R\theta$ or $p = R\rho\theta$, in which R is the gas constant equal to 2,870 when the pressure is given in kilobars or *c.g.s* units; θ equals temperature in degrees Absolute; and ρ equals the density or $1/v$. Different values of the gas constant R have been given by Shaw computed from the formula

$$\frac{R_w}{R_o} = \frac{p_w}{p_o - \frac{3\theta_o}{8\theta_o} p_w}$$

where R_o is the value for dry air; R_w for mixture of air and

water vapor which is saturated at temperature θ and has a partial pressure of dry air of 1,000 kilobars at the freezing point.

Avogadro's law

The characteristic equation is true for an ideal gas only. Another law, due to Amedeo Avogadro, states that under the same conditions of temperature and pressure equal volumes of gases contain the same number of molecules. The number of molecules in one cubic centimeter of any gas at a pressure of 1,000,000 dynes and a temperature of 273° A. is 27 billion billion.

16. Dynamical heating and cooling of the air. When air or any other gas — or, even a solid — is compressed, there is an increase of temperature which is generally apparent if the compression is sudden. Thus with a hand-compression pump: when the piston is forced down, the molecules of the gas have their kinetic energy increased; and conversely, when the gas or air is allowed to expand, the kinetic energy of the molecules is decreased, and energy has been expended in pushing back the piston. In the former case the temperature rises; in the latter the temperature falls, and it appears that the average kinetic energy is proportional to the absolute temperature. The fact that "the average kinetic energy of the molecules is equal to a constant multiplied by the absolute temperature is," says Ames in his *Constitution of Matter*, "believed to be also true for solids and liquids, the constant being the same as for a gas."

Compression increases kinetic energy

When a body falls rapidly, or slowly, or in any way moves from a higher to a lower level, work is done and resistance overcome. Potential energy due to elevation has been changed into kinetic energy expended in giving momentum to the mass. If the resistance or friction be marked, then the rise in temperature is generally noticeable. Conversely, in raising or lifting a body, work is done against the force of gravity, and potential energy is acquired at the expense of kinetic energy, or heat. Thus when a mass of air rises, as it will when heated, or falls, as it will when cooled, work is performed; and energy of either potential or kinetic kind is transformed

Kinetic energy transformed into heat

into heat. But the sum total of the energy remains the same according to the principle of the conservation of energy.

The lifting of one kilogram of dry air against the force of gravity through a distance of one meter has been taken as the unit of work, and is known as the kilogram-meter. In the *c.g.s.* system the unit would be more properly one gram lifted a distance of one centimeter, or the erg. If we take the dyne as the unit of force, then one dyne per square centimeter acting through one cubic centimeter does one erg of work.

The megerg

A million of the small units of work is called a *megerg*, although some writers prefer the term "megalerg." This causes some confusion; likewise the fact that there are two units of heat in general use: a large one much used by engineers, which is the amount of heat required to raise the temperature of a *kilogram* of water one degree; and the small unit or therm, which will raise the temperature of a *gram* of pure water one degree. Since there is a slight difference in the specific heat of water at different temperatures the small calorie, or therm, is more precisely defined as the amount of heat that will raise the temperature of a gram of pure water from 273°A. to 274°A. Now, it has been shown experimentally that the energy which can raise the temperature of a gram of water one degree could do the mechanical work of lifting, against the force of gravity, one gram of water 42,683 centimeters. This is called the *mechanical equivalent of heat* under standard gravity; that is, at 45° latitude and at sea level. This is usually expressed $1/A$. Conversely there is a *heat equivalent of work* usually expressed by the symbol A. Its value is 0.00002343.

The mechanical equivalent of heat

When heat is added to a given volume of air, say one cubic centimeter, and this is free to expand in one direction, some of the heat goes toward increasing the temperature and some is expended in the work of expansion. Expressing it more directly in percentages, we may say that of a given quantity of heat, 71 per cent is utilized in raising the temperature and 29 per cent is spent in expanding the air against atmospheric pressure. This ratio is determined as follows: To raise the temperature of a gram of pure water one degree needs one

small calorie; to raise the temperature of a gram of dry air, 0.2375 calorie, which is the specific heat of air at constant pressure. This is sometimes written

$$C_p = C_v + AR.$$

Now, a cubic centimeter of pure, dry air weighs somewhat more than a thousandth of a gram, or 0.00129305 gram; and if we multiply this by the specific heat of air we have 0.000307, which is heat that would raise the temperature one degree and expand the air 1/273 of its volume. The expansion, however, has been done against the pressure of a sea-level atmosphere, or 1,033,300 gram-centimeters. If we use the new atmosphere, we have 1,000,000 dynes divided by 273, or 3,663 dynes. This is accomplished by 0.307 unit of heat, and therefore a whole unit would do 11,931 dynes. To this we must add the work done in rising from sea level to the new base, which is 106 meters, or 348 feet, above sea level. There would also be a slight change in gravity. The total would be 12,330 dynes. But the work or mechanical equivalent of one unit of heat is 42,683 gram-centimeters, and the ratio of this to the former is as 1 to 0.29.

Energy used in expansion

When there is no expansion of air and no work is done, the whole amount of heat is consumed in raising the temperature. In this case the increase in temperature is to the increase when work is performed as 1 to 0.71, or as 1.41 to 1. In the same way, the heat that will raise the temperature of a cubic centimeter of dry air without change in volume is to the heat required when there is change in volume as 1 to 1.41.

The specific heat of air under constant pressure being 0.237, the specific heat under constant volume is 0.237/1.41, or 0.169. The difference between the specific heats is 0.068.

It follows from what precedes that where no heat is added to or subtracted from a mass of air, and it expands or contracts under varying pressure, such work must be done at the expense of its own heat. Hence the heat loss (when air expands by coming under diminished pressure adiabatically or without the addition of heat from other source) for every 1/273 part of volume increase is 0.29 of a calorie. To cool a whole degree the expansion would have to amount to 1/79

of the volume. We have seen that the height of the homogeneous atmosphere is 7991 meters; hence 7991/79 gives us approximately the height 101.2 meters, to which air must be lifted to cool one degree, or practically one degree per hundred meters. Strictly, the adiabatic rate is 9.8 per kilometer, but this value is true only for small changes.

Adiabatic condition rare

The air is not dry, but, on the contrary, especially in the lower levels, is often saturated, and so a strictly adiabatic condition is rarely met. The specific heat under constant pressure is not a constant but a variable quantity. Circulation and radiation materially alter the heat distribution. In problems of dynamic meteorology, observations which contain the height z, the pressure p, and temperature θ, but omit the velocity of circulation and the direction of motion q, are defective; and if, furthermore, the amount of water vapor present and the various changes of state are not considered, then there can be no precise data regarding heat distribution.[1]

ACTUAL TEMPERATURE GRADIENTS

The following tables, given by Dines,[2] give the fall of temperature per kilometer, or the approximate temperature gradient, for every month:

km.	0–1	1–2	2–3	3–4	4–5	5–6	6–7	7–8
Jan.	5°A.	4	4	6	7	7	6	7
Feb.	5°A.	5	4	6	7	6	7	7
March	4°A.	6	4	6	7	6	7	7
April	6°A.	6	5	6	7	6	7	7
May	6°A.	6	5	6	6	7	7	6
June	6°A.	6	5	6	6	7	7	7
July	6°A.	5	5	6	6	6	8	6
Aug.	6°A.	4	5	6	6	7	7	7
Sept.	5°A.	3	5	6	6	7	7	6
Oct.	4°A.	5	5	6	6	7	6	7
Nov.	5°A.	3	5	6	6	6	8	6
Dec.	5°A.	3	5	6	6	7	7	6
Average	5.3	4.8	4.8	6.0	6.3	6.6	7.0	6.6

[1] F. H. Bigelow, "Thermodynamics of the Earth's Non-adiabatic System," *Am. Jour. of Sci.*, Dec., 1912. Also same author's "Treatise on Circulation and Radiation in the Atmospheres of the Earth and of the Sun," 1915.

[2] *Philos. Trans. of the Royal Soc. of London*, Series A, Vol. 211, pp. 253–278.

km.	8–9	9–10	10–11	11–12	12–13	13–14	Mean 0–9
Jan.	6	4	3	0	1	0	5.8
Feb.	6	3	3	−1	1	0	5.9
March	6	4	3	−2	0	0	5.9
April	6	4	3	−1	−1	0	6.2
May	7	5	4	−1	−1	0	6.2
June	7	6	4	−1	−1	0	6.3
July	7	8	4	0	−1	1	6.1
Aug.	8	7	4	1	0	0	6.2
Sept.	8	7	5	1	1	0	5.9
Oct.	7	7	4	1	1	1	5.9
Nov.	7	5	4	1	1	1	5.8
Dec.	7	4	3	1	1	1	5.8
Average	6.8	5.3	3.5	−0.1	0.2	0.3	6.1

Gradient per thousand meters

Thus it is evident that the gradient in the free air under usual conditions is approximately 6° per thousand meters, while under adiabatic conditions the fall is 9.8°.

NOTE.—The student will find detailed free-air data in Supplement No. 3 of the *Monthly Weather Review* issued Dec. 1, 1916.

CHAPTER V

Stratosphere and Troposphere

17. The stratosphere and troposphere. The most important outcome of the numerous soundings of the upper air has been the discovery of a cessation of fall in temperature at a certain height. Above this level the temperature remains stationary or even rises, and it has been found that the height at which the gradient ceases or reverses varies with season and latitude.

Gradient ceases at a certain height

The actual cessation of the fall in temperature was first noticed by Teisserenc de Bort in June, 1899, and confirmed in March, 1902. The phenomenon was discussed in May, 1902, by Assmann, who made special experiments to prove that the condition was not the result of defective ventilation of the thermometers employed. The absence of a vertical decrease of temperature led to the use of the name *isothermal layer*, or upper inversion; but de Bort soon introduced the terms *stratosphere* for the upper layer, and *troposphere* for the lower levels where convection did occur. Gold suggested the term "advective" for the upper region; because any interchange of air would occur chiefly through horizontal motion, while in the lower the interchange would take place through ascensional and descensional currents, and therefore might well be called "convective." These names, however, have not come into general use. There are frequent inversions of temperature near the earth's surface and it has been suggested by the author that these be called the minor inversions, while the change at the stratosphere be called the major inversion.

The upper inversion

Major and minor inversions

The average height at which the change occurred was found by Teisserenc de Bort to be 11 kilometers; and he also called attention to the fact that the height varied in highs and lows, averaging 12.5 kilometers in the former and 10 kilometers in the latter. At a

Variation of average height

level of 10 kilometers the difference in pressure between an average high and an average low would be approximately 10 kilobars, while the difference between the pressures at the level of the stratosphere in ordinary highs and lows is about 70 kilobars.

Record of soundings made at Uccle

Numerous records of soundings have been made public through the International Commission for Scientific Aëronautics (see p. 13). A good illustration of a successful sounding is given in abridged form below. The ascent was made at the observatory at Uccle, Belgium, June 9, 1911, during pleasant weather. A height of 31,780 dynamic meters, or 32,430 meters, was reached. A temperature of 212°A. was recorded

Time	Pressure kb.	Pressure mm.	Elevation Met.	Temperature Abs.	Gradient Δt/100m.	R. H.
7:00	1,001	751	100	290°	....	81
7:05 (?)	900	675	1,000	287	....	..
....	797	598	2,000	281	.056	..
....	705	529	3,000	275	.066	51
7:18	621	466	4,000	269	.042	34
7:20 (?)	547	410	5,000	264	.069	..
....	479	359	6,000	251	.086	30
7:34	313	235	9,000	234	.084	30
7:38	271	203	10,900	222	.065	..
7:44	199	149	12,000	212	.039	29
7:47:04[1]	168	126	13,040	213	.021	29
7:48	160	120	13,340	213	−.007	29
7:52	129	97	14,650	218	−.031	30
7:56	103	78	16,050	223	−.07	29
8:02	72	54	18,370	218	.01	29
8:12	36	27	22,720	222	.00	29
8:32	8	6	32,430	234	....	29

[1] Beginning of inversion.

at the 12,000-meter level. It may be remembered that the lowest temperature experienced by Scott was about the same.

A very comprehensive review of the results of various ascents made under the auspices of the International Commission is given by G. Nadler in the *Beiträge zur Physik der freien Atmosphäre*, VI Band, Heft 2, issued December, 1913.

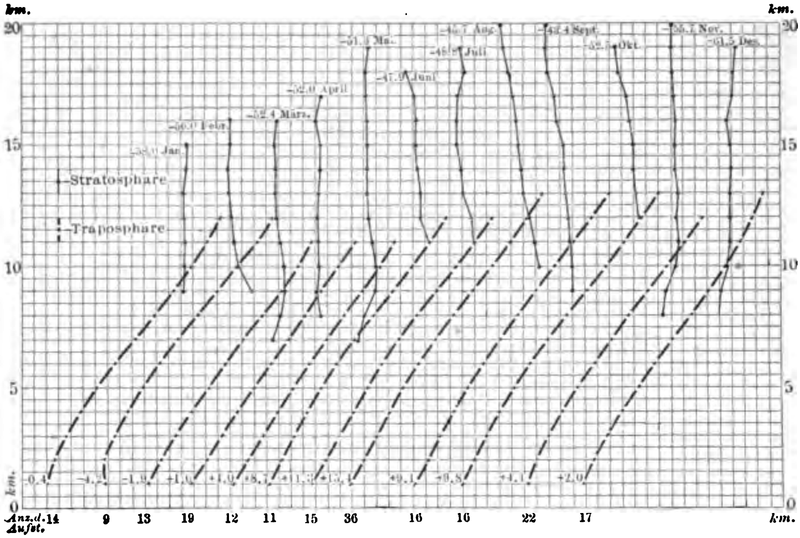


Fig. 10. Monthly Values of Temperatures

The figures beneath the diagram (*Anz. d.* and *Aufst.*) indicate the number of soundings upon which each monthly curve is based. The figures at the right and left express the altitude in kilometers, and those on the diagram itself, the temperatures in degrees C. For example, 15 sounding-balloon ascensions in July gave an average temperature of 284°A. (+11° C.) at altitude one kilometer. The temperature decreased up to altitude 12 kilometers and then remained almost constant, or increased slightly to altitude 19 kilometers, temperature 224°A. (–49° C.).

The values doubly underlined in the table on the following page indicate the lowest temperature reached, which practically is the beginning of the isothermal column. The commencement of this is generally indicated by the symbol H_c, and the temperature at the bottom T_c. The figures with single underlining show the freezing temperature, which descends to sea level in latitudes 62° north and south, but which, at the equator, is more than 5 kilometers above sea level.

Variation of stratosphere with latitude

Above the equator the regular fall in temperature with elevation continues to a much greater height than in temperate regions. Furthermore, the lowest temperatures in the upper air are found above the equator. This was indicated by de Bort and Rotch as early as 1905 in their exploration of the air over the north tropical Atlantic. They also demonstrated that above the trade winds there was a change of wind direction; that is, above the

northeast trade to southwest and above the southeast trade to northwest, the height of the plane of reversal varying greatly, while near the equator the wind was east at all heights. In 1908 an expedition from the Lindenberg Observatory sent up balloons from Victoria Nyanza, on the equator, and two of these reached the stratosphere, giving values of 17.2 and 15.4 kilometers for H_c, and 190°A. and 203°A. for T_c. But the best confirmation of this variation with latitude is found in the records from Batavia, Java, made on December 4, 1913, when a temperature of 182°A. was recorded. The balloon reached a height of 26 kilometers, but from 17 kilometers (the H_c) upward the temperature rose steadily. In six years, 1910–1915, Dr. van Bemmelen obtained sixty-six records out of 103 ascents. The mean height of the stratosphere is just

Kilometers	Jan.	Feb.	Mar.	Apr	May	June	July	Aug.	Sept.	Oct.	Nov	Dec.
14	216°A	17	19	21	22	23	22	21	19	17	16	215°A.
13	16	17	19	21	22	23	22	21	19	18	17	16
12	17	18	19	20	21	22	22	21	21	19	18	17
11	17	17	17	19	20	21	22	22	21	20	19	18
10	20	20	20	22	24	25	26	26	26	24	23	21
9	24	23	24	26	29	31	34	33	33	31	28	25
8	30	29	30	32	36	38	41	41	41	38	35	32
7	37	36	37	39	42	45	47	48	47	45	41	38
6	43	43	44	46	49	52	55	55	54	51	49	45
5	50	49	50	52	56	59	61	62	61	58	55	52
4	57	56	57	59	62	65	67	68	67	64	61	58
3	63	62	63	65	68	71	73	74	73	70	67	64
2	67	66	67	70	73	76	78	79	78	75	72	69
1	71	71	73	76	79	82	83	83	81	79	75	72
Ground	276	76	77	82	85	88	89	89	86	83	80	277

under 17 kilometers. At an elevation of 4 kilometers the temperature is 273°A., at 10 kilometers 239°A., at 17 kilometers 189°A., this latter value based on twenty observations.

The following table (p. 50) of mean heights and temperatures at the base of the stratosphere is given by Gold.[1]

The chart on p. 51 (Fig. 11) drawn by Professor Rotch, and somewhat modified from the original, shows the height of the isothermal stratum at various latitudes and corresponding heights of the isotherm of 273°A.

[1] *Geophysical Memoirs*, 5, p. 110.

Two remarkable features of the stratosphere are, then, the great elevation in tropical regions and the lowered temperature.

Variation of stratosphere with season The height of the stratosphere also varies with the seasons and with pressure distribution. Occasionally large variations are found in T_c and H from one day to another. Thus Gold quotes the conditions on April 1, 2, and 3, 1908, when the values were: H_c, 10.5 km., 12.0 km., and 7 km.; T_c, 215°A., 217°A., and 224°A.

MEAN HEIGHT AND TEMPERATURE OF THE BASE OF THE STRATOSPHERE

	H (km.)	T (°A.)
Berlin	10.34	215
Munich	10.73	216
Strassburg	10.72	215
Vienna	10.33	216
Hamburg	10.13	218
Paris	10.55	217
Uccle	10.93	213
Zürich	10.2	218
Koutchino	10.5	215
Pavlovsk	9.6	219
England	10.6	217
Italy	11.0	214
Equator	17.0	182

Various explanations for the existence of the stratosphere have been offered by Trabert, Fényi, Gold, Humphreys, Emden, Braak, and others. In general, the existence of the stratosphere is explained as due to radiation, Gold finding that above the isobaric level of 250 kilobars radiation has a heating effect, and below this, a cooling effect. This hypothesis is based on the assumption that convective temperature equilibrium exists, and on the fact that there is the usual decrease of water vapor with elevation. The actual temperature of the stratosphere, as theoretically determined by Gold, is:

Stratosphere due to radiation

$$203°\text{A.} + \tfrac{1}{4}(240° - 203°) = 212°\text{A.}$$

The theory indicates that if convection were absent and the absorption of solar radiation did not increase with height, the normal state would be one in which the gradient of temperature diminished gradually to a very small value. Emden derives

Minimum radiation temperature

a minimum radiation temperature of 214°A. by making allowance for a small quantity of water vapor. Braak holds that the very low temperatures observed in the upper part of the tropical troposphere (which are about 30° lower than those at the same height in the temperate regions) must be connected in the low-pressure belts of the tropics with the rising air currents of the general circulation, which disturb the temperature distribution, as determined by radiation and absorption, and shift the troposphere to greater heights. The fact that at Batavia the upper limit of the anti-trade winds is at the same height as the base of the stratosphere proves that convection currents reach as high as the upper limit of the troposphere. Braak says, "The upheaval of the stratosphere in the tropics may be demonstrated in a very instructive way by comparison of its height with the height of the cirrus clouds." The base of the cirrus, in his opinion, represents fairly well the height of the hypothetical dividing surface between the cooling and heating effect of radiation for moist air. The surface is one of fairly uniform temperature, as shown by the temperatures at the cirrus level in the following table:

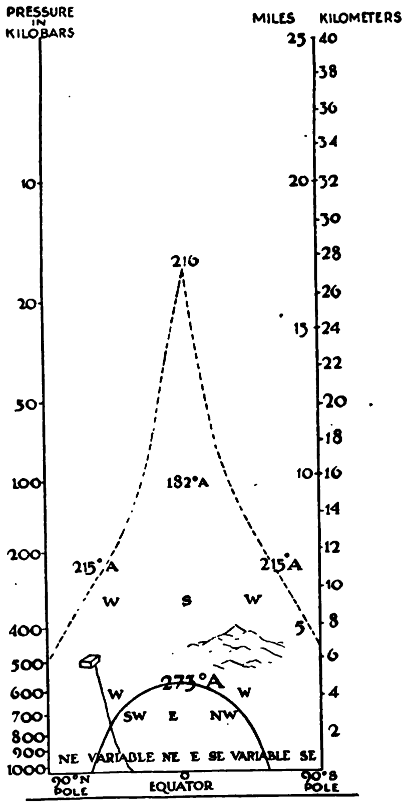
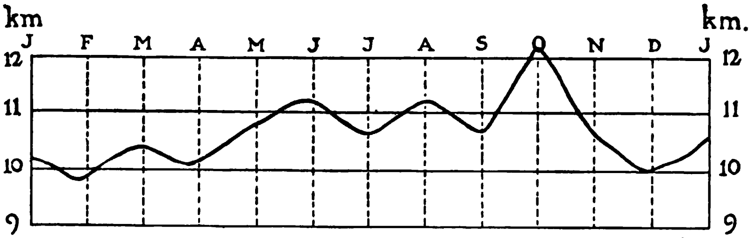


FIG. 11. ROTCH'S DIAGRAM OF HEIGHT OF STRATOSPHERE WITH LATITUDE

Bossekop, 70° N. lat., height of cirrus 8.3 km., temp. 228°A.
Potsdam, 52° N. lat., height of cirrus 9.2 km., temp. 227°A.
Batavia, 6° S. lat., height of cirrus 11.4 km., temp. 225°A.

At Blue Hill, 42° N. lat., the average height of cirrus is 8.9 km. Parallel to this surface or base of the cirrus is the base of the stratosphere, with a nearly constant temperature of 218°A.



After Gold

Fig. 12. Annual Variation in Height of Stratosphere

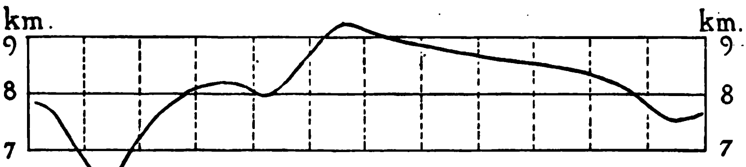


Fig. 13. Height of Cirrus Clouds

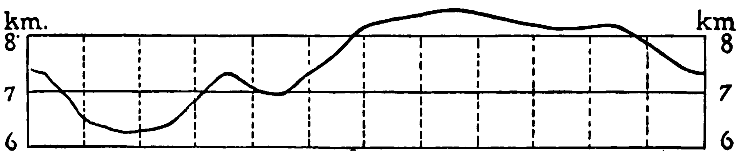


Fig. 14. Height of Cirro-Stratus Clouds

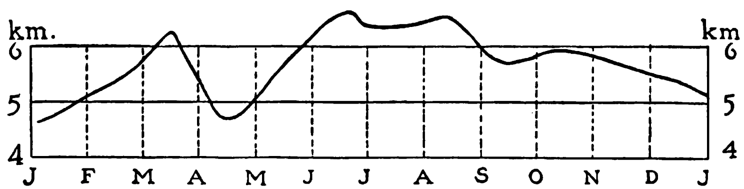


Fig. 15. Height of Cirro-Cumulus Clouds

Braak further points out that there is an essential difference between the cloud formation in the cirrus level and above it, as compared with the lower regions, a phenomenon very evident in the quiet tropical atmosphere. The upper part of the high cumulus clouds (estimated at 13 or 14 kilometers) does not, like the lower part, dissolve rapidly, but, assuming a flattened form and cirro-stratus-like appearance, it drifts along for a

Behavior of cumulus cloud near stratosphere

considerable time. This difference he attributes to a cloud-dissipating (cooling) effect of radiation in the lower, and a cloud-forming (heating) effect in the upper, levels. He believes that the lower limit of the cirrus clouds may be regarded as the level where, for air of abundant water content, the influence of radiation changes its sign.

Gold gives extensive comparisons of the various upper clouds which may show the same peculiarities as the annual variation in the value of H_c (Figs. 12–15). The actual heights, however, are much less for the clouds, and he is of opinion that, while the results point to some common cause, they indicate that the formation of clouds is not a usual cause of the sudden decrease and change of sign in the temperature gradient.

CHAPTER VI

The Circulation of the Atmosphere

18. Effect of the earth's rotation on the atmosphere. If the earth were at rest, the general motions of the atmosphere would be materially different from what they are. Furthermore, certain equations of motion which are true for small and level areas do not hold for motions on a larger scale covering a considerable area of the earth's surface, that is, the surface of a rotating sphere. It may be pointed out that, on a stationary earth, warm, light air would ascend and cold, heavy air descend, and thus the pressure would be greatest at the poles and least at the equator. This is not the case on the rotating earth, where there are great polar depressions. Moreover, we find that, in general, the ascending air of cyclones is cold and the descending air of anticyclones relatively warm and comparatively light.

Terrestrial rotation and air movements

Our knowledge of the circulation of the atmosphere as a whole is still very imperfect. While the early navigators doubtless knew of the existence of certain well-marked winds, there was no definite knowledge of even such steady winds as the trades and monsoons until about the end of the seventeenth century.

The earliest knowledge of steady winds

Edmund Halley, who discovered the comet which bears his name, was one of the ablest physicists of his age and was a friend of Newton (it was through Halley's efforts that the *Principia* was published). Moreover, he was an explorer and a navigator. He was the first[1] to attempt a magnetic survey (1700) of the ocean, and it is in connection with his charts showing variation of the compass that we find advanced a theory of the general and coasting trade winds and monsoons or shifting trade winds. And it may not be out of place, as illustrating the constancy of the great air streams, to quote a

First definite knowledge of trades and monsoons

[1] Gellibrand, in 1634, definitely proved that the compass direction varies from year to year.

passage from a lecture[1] by Bauer showing that if the magnetic survey vessel, the *Carnegie*, which has circumnavigated the globe and repeatedly intersected the course of the *Paramour Pink*, the sailing vessel which Halley commanded, were to set course from "St. Johns, Newfoundland, and follow the same magnetic courses as those of the *Paramour Pink*, instead of coming to anchor in Falmouth Harbor she would have made a landfall somewhere on the northwest coast of Scotland. In brief, *while the sailing directions as governed by the winds over the Atlantic Ocean are the same now as they were during Halley's time*, the magnetic directions or bearings of the compass that a vessel must follow to reach a given port have greatly altered."

Constancy of the great air streams

Halley's voyage

In 1735 another English astronomer, Hadley, advanced an explanation of the trade winds based upon a change of velocity in air moving north or south because of change in the rotational velocity of the earth with distance from the pole. Air moving from high latitudes southward would pass to regions of greater velocity and hence would lag or apparently blow westward. This view held for many years; but finally a flaw was found in the reasoning. Air moving relative to the earth will, because of inertia, maintain its absolute velocity but will change its relative velocity, which is the resultant of the absolute velocity and the reversed velocity of the point of reference.

Hadley's theory of the trades

Different points on the earth's surface have different absolute velocities on account of the earth's rotation, and hence air would change its relative velocity in moving north or south; but, as will be shown later, the deflecting force acts at right angles to the direction of motion. G. Coriolis, in 1835, gave the first mathematical solution of this problem of the earth's deflective effect; but the first[2] one to apply this deflection to air motion was William Ferrel in 1859, in his general discussion of the motion of fluids relative to the earth's surface. Ferrel showed that the natural deviation due to the earth's

The earth's deflective effect

[1] The fourth Halley lecture, delivered at Oxford, May 22, 1913; reprinted in the Smithsonian Report, 1913, entitled "The Earth's Magnetism."

[2] See also Charles Tracy in *Am. Jour. Sci.*, 1843.

rotation could be counterbalanced by pressure distribution where the gradient was such as to cause an acceleration sufficient to offset the effect of rotation. The deflecting force due to the earth's rotation acts at right angles to the direction of motion. Ferrel's scheme of planetary circulation as later developed does not, however, meet modern views. It requires a rather involved flow of air from the equator to the poles, also in the higher levels; and there is no special evidence supporting such a theory of this upper circulation. The circulation, as outlined, requires marked depressions around the poles, whereas, in reality, an entirely different distribution of pressure exists.

Ferrel's scheme of planetary circulation

Ferrel gives an interesting illustration of the effect of a deflecting force by the experience of a person walking over a narrow drawbridge while it is turning on its pivot. If there were no railings the tendency would be to go over the side; and if there were railings, to be forced against them. This tendency would be in proportion to the velocity of transit across the bridge and the angular velocity of its gyration. Somewhat similar is the deflective effect of the earth's rotation. A body in motion in any direction relative to the earth's surface tends, if free to move, from this direction. This deflecting force moves the air to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. The deflecting force is usually expressed by the formula

Illustration of effect of a deflective force

Value of deflective force

$$2m\omega V \sin \phi,$$

where

$$\omega = \text{angular velocity of earth's rotation}$$

$$= \frac{2\pi}{86{,}164 \text{ sec.}} = 0.00007292;$$

V, the velocity with which the body is moving relative to the earth; ϕ, the latitude; and m, the mass.

Briefly stated, moving air is deflected to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. Components of gravity cause the deflections. For

east and west motions the direction but not the velocity is changed; while for north and south motion, the velocity is affected. Air moving toward the pole has its eastward velocity increased, and air moving toward the equator has its velocity diminished. It will be well to recall the law of equal areas or constancy of angular momentum; also the law of constancy of density multiplied by velocity, sometimes called Egnell's law, or Clayton's law.

Deflection of moving air in general

If there is no velocity,— that is, if it is calm and there is no air movement,— there is no acceleration. Again, at the equator the sin ϕ is zero. The factor ω, or the earth's angular velocity, is found by dividing the circumference by the number of seconds in one sidereal day. It may also be pointed out here that acceleration due to centripetal force must be deducted from gravity, and the remaining component of gravity, called apparent gravity, or gravity as ordinarily measured, is somewhat less than absolute gravity, as calculated by astronomers; and it acts in a slightly different direction.[1]

Application of formula for deflecting force

Ekholm sums up[2] the chief deviations due to this deflective effect as follows:

"In the Northern Hemisphere in an obliquely upward current the horizontal component of the deviating force of a south wind is less, and that of a north wind is generally more, than in a horizontal current. In an obliquely downward current the reverse holds good. As to the Southern Hemisphere, we need only to interchange the north and south in the above proposition. . . . The north and south winds are never *vertically* deviated.

Ekholm's summary of deviations due to deflective effect

"In an east-west current, if the air is obliquely rising or falling, the deflective force will not be directed exactly perpendicular to the velocity; but if the air is rising it throws the east winds a little forward and the west winds a little backward; and the reverse of this, if the air is falling. The west

[1] A geometric method of deriving the deflective force due to the earth's rotation is given by T. Okada in *Monthly Weather Review*, May, 1908, p. 147; also by C. F. Marvin, *ibid.*, October, 1915, p. 503.

[2] *Monthly Weather Review*, June, 1914, p. 330.

winds always deviate upward and the east winds downward."

The following example may illustrate the deviation in absolute units:

"If the air temperature is 290°A. and the pressure 1,000 kilobars, and the density of the air equal to 0.0012, the acceleration will be equal to the gradient g; that is, the acceleration produced by g will be expressed by the same measure of length as the gradient itself, if this is given for a meridian degree. Now we know by observation the relation between gradient and wind velocity in a steady motion. In a cyclone, for example, a gradient of 5 millimeters will generally produce stormy winds with a velocity of about 30 meters per second ($v = 3,000$ cm./sec.). Then, since $\omega = 0.00007292$, we get for a current parallel to the equator a deflecting force equal to 0.44 cm./sec.², which is nearly the same value as that for the acceleration produced by the gradient, this latter being 0.5 cm./sec.²" The discussion is carried further by Ekholm, not omitting the effect of friction (the coefficient of friction, k, ranges from 0.00002 to 0.00004 for the open sea, and does not exceed 0.00012 for a very rough continental surface; and the friction would therefore be kv, or less than the deflecting force in a current parallel to the equator). "Friction in a horizontal parallel current is so small in the upper strata that it is negligible compared with the gradient. Hence it follows that in the upper strata even a very slight gradient,—for example, 0.01 centimeter,—if it act sufficiently long in the direction of motion, can produce a marked velocity, particularly as the acceleration is inversely proportional to the density of the air. Thus at an altitude where the density is only half that at sea level (about 8 kilometers, or the average height of cirrus) the acceleration produced by the above-named slight gradient will be 0.03 cm./sec.² This acceleration acting in the direction of motion during 100,000 seconds or 28 hours will cause a velocity of 3,000 cm./sec. or 30 meters per second; and with this velocity the deflective force would be 0.44 cm./sec.², or about 15 times greater than the component of acceleration acting in the direction of motion.

Relation between gradient and wind velocity

Effect of friction on deviation

Acceleration produced by a slight gradient

"The horizontal component of the deviating force attains its greatest value in the higher latitudes; and the vertical component in lower latitudes."

19. Impressions of gyroscopic motion. Sandström makes the pertinent suggestion[1] that much of the difficulty experienced in comprehending gyroscopic motion is due primarily to the absence of any specific sense that would enable us to detect, if not to feel, the earth's rotational effects. He says: "From childhood on, we are accustomed to regard the visible portion of the earth's surface as at rest. To be sure, we all really know that the earth does rotate and we can imagine this, but we have no sense by which to feel it. . . . The constant deception that the earth is at rest impresses the observer as being the true state of affairs. Under these conditions it is indeed very natural that even the effects of the earth's rotation should appear foreign to us." And again: "A being who could feel the terrestrial rotation would probably find it very natural that specifically heavy air should ascend, and light air descend. That person would find it easy to apply Coriolis' theorem, because the resulting conclusions would be in harmony with his sensations."

Gyroscopic motion difficult to comprehend

The best means of bringing home the effects of rotation upon the air and water of the earth are hydrodynamic experiments with rotating vessels. Sandström considers these experiments from two hypothetical standpoints. The first would be that in which a very small individual was on the rotating vessel, a being so small and with such limited observational powers that he could not recognize the rotation of the vessel. To such a person many of the processes would appear inexplicable. For example, by experiments carried out on a small scale he would find that liquids and gases specifically lighter than their surroundings will strive upward, but would find just the opposite in experiments on a large scale. Eventually, by numerous experiments he might come to the conclusion that the vessel had a rotating motion.

Hydrodynamical experiments with a rotating vessel

Experiments considered from the relative point of view

[1] *Monthly Weather Review* Sept 1914, p. 523.

With mathematical analysis he might work out a theorem, such as Ferrel's or Coriolis'.

The second viewpoint would be that in which the observer saw the rotating vessel and the rotary movements. This second method would present matters as they would appear to an observer outside of the earth. Studying oceanic and atmospheric circulations, he would see a series of large vorticular motions and could comprehend the effects of centrifugal forces. Sandström emphasizes the advantage of this second method, since "one may thereby see and judge of the whole absolute motion and the associated forces without any intermediary. In the former method one is concerned with two motions: the relative motions of the atmosphere or of the ocean water as referred to the earth's surface, and the motions of the earth's surface as the result of its rotation. The compounding of these two motions and of the forces that bring them into existence is not always an easy problem."

Experiments from the absolute point of view

He suggests that the experiments, as outlined below, be considered first from the absolute (that is, the second) point of view and then be transferred to the relative (first) point of view. In this way he thinks that we may gradually acquire the ability intuitively to take immediate account of the effect of terrestrial rotation when discussing the observed movements; and he considers this acquirement as one of the most important objects of dynamic meteorology.

The heaping of ocean water explained

To explain the heaping up of warm ocean water in the horse latitudes,[1] he fills a glass vessel 30×10×10 centimeters to a depth of about 3 centimeters with fresh water, and introduces gently below the fresh water an equal volume of salt water. One of the strata is colored for purposes of better observation. By means of a bellows, a tube, and the perforated spout of a watering pot, a current of air is forced downward upon the water surface as in Fig. 16. At once it becomes

[1] This name has its origin in the fact that in early days vessels bound from New England to the West Indies, carrying horses, were so often delayed by calms that for want of water it was necessary to throw the horses overboard. These calms are experienced between the prevailing westerly winds north of 25° N. latitude and the northeast trade winds.

clear that the bounding surface between the two strata bulges upward, a result of the downward air current driving the surface water toward the sides of the vessel. If the vessel is now placed on a rotating table, and a slow rotation around a vertical axis begun, it will be found on using the bellows as before that the bounding surface of the two layers does not bulge upward at a point beneath the spout, but, on the contrary, is depressed, as shown in Fig. 17. From the experiment one may conclude that the heaping of ocean water in the horse latitudes is a result of the earth's rotation in combination with the anticyclonic conditions prevailing there.

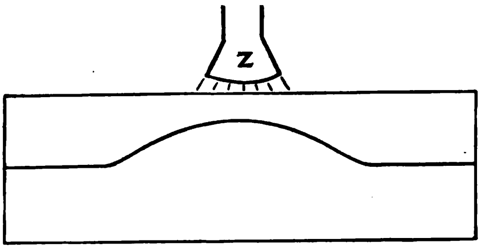


FIG. 16. EFFECT OF RADIALLY DIRECTED WINDS UPON A SYSTEM AT REST

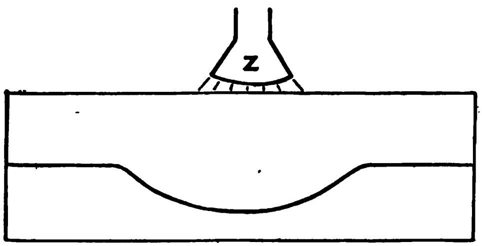


FIG. 17. EFFECT OF RADIALLY DIRECTED WINDS UPON A ROTATING SYSTEM

Sandström then endeavors to explain the phenomenon from the viewpoint of the small imaginary being referred to above, who can observe only the air movement relative to the rotating vessel and perceives only that the air blows spirally outward from a center (Fig. 18). This air movement produces also an anticyclonic circulation in the water, whereby, on account of the deflective force, this rotation comes into action and drives the water toward the right hand. In other words, it presses toward the center, and the water heaps up at the center. From the other viewpoint it is seen at once that the vessel is rotating and that a stream of air is blowing radially from a central point upon the water surface (Fig. 19). The lower stratum has the same velocity of rotation as the vessel itself, but the rotation of the surface water is hindered by the

Anticyclonic circulation

radial currents, and the surface water moves more slowly than the vessel. The centrifugal force of the lower stratum is greater than that of the upper. Hence the lower water is forced outward while the upper stratum collects in the center.

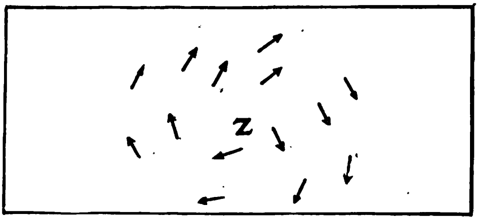


FIG. 18. RELATIVE MOTION OF RADIALLY DIRECTED WINDS AT THE SURFACE OF A ROTATING SYSTEM (VIEWED FROM ABOVE)

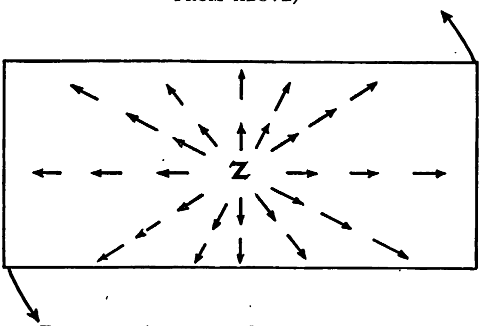


FIG. 19. ABSOLUTE MOTION OF RADIALLY DIRECTED WINDS AT THE SURFACE OF A ROTATING SYSTEM (VIEWED FROM ABOVE)

In consequence of the earth's turning on its axis its surface is everywhere in cyclonic rotation and the velocity of rotation is determinable by the Foucault pendulum experiment. If the earth's angular velocity of rotation is ω, then the angular velocity of any point on the earth's surface is $\omega \sin \phi$, where ϕ is the geographic latitude. The time required by the earth to complete one rotation is 24 hours. *Air, then, that is apparently at rest, has a cyclonic circulation; in fact, even the air in the familiar anticyclonic whirl actually possesses a cyclonic rotation.* Air, then, apparently at rest, is subject to a centrifugal force which is reinforced under cyclonic conditions and weakened under anticyclonic circulation. These facts explain the temperature distribution within cyclones and anticyclones. In a cyclone the rotation at first increases, with altitude reaching a maximum at a certain height. The centrifugal force is greatest at that level, and the air is driven most strongly outward. The air therefore undergoes dynamic cooling below the level of maximum cyclonic rotation and dynamic warming above that level.

Cyclonic movement

Velocity increases with elevation

In an anticyclone the anticyclonic rotation has its maximum at a certain level where the centrifugal force is lowest, and from that level the centrifugal force increases both upward and downward. Consequently, below this level the air will be drawn downward, and above, it will be drawn upward; therefore, below this level the air will be warmed dynamically, and above, it will be cooled.

West-east drift of the atmosphere

Again, the west-east drift of the atmosphere in middle and higher latitudes forms a gigantic polar cyclone. This west-east drift has its maximum at a certain level and diminishes both upward and downward therefrom. At the level of the maximum drift the centrifugal force is the greatest; below that level the air of the polar regions is drawn upward, and above it the air is drawn downward; therefore, beneath this level the temperature of the air at the poles is lower than it is at the equator, while above this level the air is warmer above the region of the poles than it is at the same level over the equator.

Buys-Ballot's law

In 1853 Coffin, studying the winds of the globe, noted that on the left of prevailing winds the pressure was low; and in 1857 Buys-Ballot discovered the law that if you stand "with your back to the wind, the pressure is low on the left and high on the right." In 1860 he pointed out the connection between the above relation and the rotation of the earth; and later (1862), when Buchan mapped the distribution of pressure over the surface of the globe, the close relation between wind and barometric gradient was clearly shown. Guldberg and Mohn (1876), Sprung, Weihrauch, Ekholm, W. M. Davis, and others have written on the subject. Shaw, however, was the first (1893) to point out the necessity for calculating and utilizing gradient velocities. Largely because of the experiments of Dines, it has become evident that the change in the velocity, with height,

Balance of barometric gradient by velocity

brings these upper currents more and more into agreement with the theoretical wind computed from the surface gradient. Gold, in 1908, showed that the wide separation of isobars in the inner regions of an anticyclone, as compared with the closeness of the lines in the central region of a cyclone, was dependent upon

the fact that in an anticyclone the curvature gradient acts in the opposite sense to the general rotation gradient, whereas in a cyclone the curvature gradient and the rotation gradient are concurrent. Thus, says Shaw, "the idea of the balance of barometric gradient by velocity, as a primary law of atmospheric circulation, has been gradually strengthened."

NOTE.—A schematic representation of the forces causing and modifying winds may be found in the Introductory Note, by Professor C. F. Marvin, to *Weather Forecasting in the United States*, W. B. 583, issued in 1916. On pp. 22 and 23 numerical data are given.

CHAPTER VII

THE MAJOR CIRCULATIONS

20. Hyperbars and infrabars. There are at least three causes operating in the establishment of the major circulations of the globe and less directly in the minor circulations. These are (1) the unequal heating of equatorial and polar regions (and, in the last analysis, gravity is the prime factor in causing air motion); (2) the deflective forces due to the earth's rotation; and (3) the unequal absorption of heat by land and water surfaces, which, as we shall see later, determines largely the location of the hyperbars and infrabars, or centers of action. The distribution of pressure over the continents and oceans determines in large measure the path and frequency of smaller circulations. Teisserenc de Bort, in 1881,[1] gave the name "grand centers of action" to certain areas of high and low pressure, and traced a relation between the position of these centers and periods of abnormal temperature. De Bort thus explained certain abnormal winters; a relation which had been suggested by Hoffmeyer, in 1878, and which has since been discussed, by Fassig in connection with abnormal Marches on the Atlantic coast, by McAdie in connection with winter rain on the Pacific coast, and by Humphreys on warm and cold winters in the eastern part of the United States.

Causes of the major circulations

"Grand centers of action"

Beginning with the largest ocean, the Pacific, we find two well-marked hyperbars, or areas where the pressure is in excess. One is west of California and extends southwest; and the other is over the southern Pacific, west of Chile. There are also certain continental hyperbars to which reference is made below. Over the Atlantic there are two hyperbars, a small Bermuda one and a larger one, west of southern Africa, extending from 15° E. to 30° W.

Areas where the pressure is in excess

[1] "Etude sur l'hiver de 1879–1880," *Ann. du. Bur. Cent. Météorol. de France*, IV, 1881.

and from 10°S. to 40°S. The fifth hyperbar is over the Indian Ocean. The land hyperbars are over western North America, southwestern Europe, and central Asia. There are certain peculiar reversals in these areas between summer and winter, as, for example, the North American hyperbar, which becomes an infrabar in summer; and the Australian hyperbar of July (the winter season), which becomes an infrabar in January (the summer season for those latitudes). The more prominent infrabars, or areas of diminished pressure at the surface, are the Aleutian of the North Pacific and the Icelandic of the North Atlantic. Thus, in a general way, we may, following Buchan, place the winter hyperbars in the northern hemisphere between 20° and 40°, except over land surfaces where they extend farther north. In the southern hemisphere the hyperbars are more evenly aligned, and we find them like peaks in the belt of prevailing high pressure between latitudes 20° and 40°.

Seasonal reversals

Position of winter hyperbars

Humphreys has advanced the view that hyperbars occur only where cold ocean currents cross the belts of high pressure. This explanation, however, overlooks the continental hyperbars. The general character of the season — such as the predominant wind, temperature, or rainfall — has been connected by various writers with the persistency of the hyperbars. There seems to be some correlation between abnormal weather and the shifting of these pressure areas. Hann showed that the pressure changes between the Azores hyperbar and the Icelandic infrabar were of unlike character, rising pressure in one being attended with falling pressure in the other; and falling pressure in the Iceland area causing warmer weather over central and northwestern Europe.

Relation between persistency of hyperbars and character of season

In various papers[1] published in recent years attention has

[1] Fassig, *Am. Jour. of Sci.*, Nov., 1899; McAdie, "Climatology of California," *U. S. Weather Bureau, Bulletin L*, 1903; Lockyer, *Science Progress*, Oct., 1906, No. 2; Okada, *Central Meteorological Observatory Bulletin*, No. 4, Tokyo, Japan, 1910; Humphreys, "Origin of the Permanent Ocean Highs," *Bulletin U.S. Weather Bureau*, Oct., 1911; *Monthly Weather Review*, Dec., 1914; McAdie, "Forecasting the Water Supply in California," *Monthly Weather Review*, July, 1913, also *Bulletin Mount Weather Observatory*, Dec., 1910, and *Monthly Weather Review*, April, 1908.

been directed to the relation between pressure distribution and the character of the season. On the Pacific slope typical wet winters occur when the North Pacific infrabar overlies the continent west of a line drawn from Calgary to San Francisco. Typical dry winters are associated with a westward extension of the continental hyperbar to the coast line and a retreat of the Aleutian infrabar to the northwest.

Relation between pressure distribution and character of season

In a normal season the Aleutian infrabar extends from latitude 40° N. to 60° N. and from longitude 130° W. to 140° E.

In summer months the distribution of pressure changes. The Aleutian infrabar practically disappears. The continental hyperbar is displaced somewhat eastward, and the oceanic hyperbar moves farther north. Summer in California is practically rainless, and there are strong west and northwest winds.

The accompanying charts (Figs. 20 and 21, p. 68) show the pressure distribution during selected dry and wet weather months. Fig. 20 shows the sea-level pressure and surface winds during January, 1902, typical of a dry winter month. Other dry winter months were January, 1898 and 1904, February, 1899, 1890, and 1912, and March, 1898, 1901, and 1908.

Fig. 21 shows the sea-level pressure and surface winds during February, 1902, which are typical of a wet winter month. Other wet winter months were January, 1906, 1907, 1909, 1911, February, 1904 and 1909, and March, 1904, 1907, and 1911.

The precipitation during January, 1902, a dry winter month, was so scant that the deficiency in water supply was approximately 40,000 million cubic meters. The precipitation during February, 1902, was so heavy that the excess of water supply for the month amounted to approximately 53,000 million cubic meters.

It is also to be pointed out that the frequency and path of individual disturbances depend primarily upon the strength and location of the larger areas. Individual lows move rapidly southward when the continental high overlies Idaho,

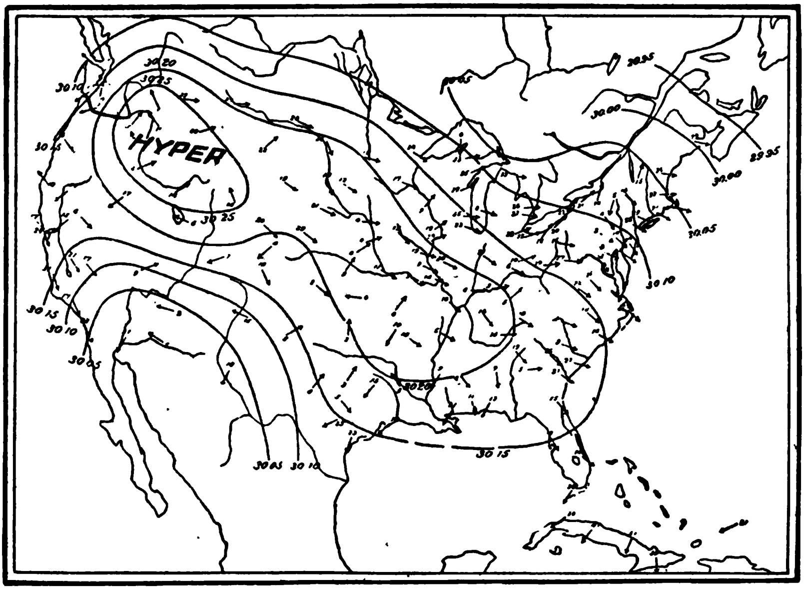


FIG. 20. TYPICAL PRESSURE DISTRIBUTION AND WIND DURING A DRY WINTER MONTH

The hyperbar, 1024 kilobars (30.25 inches), favoring north winds over California

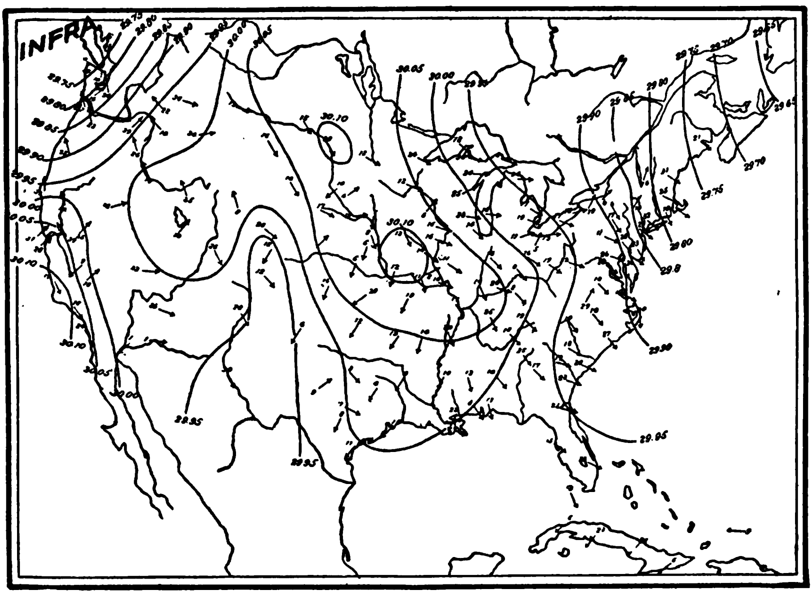


FIG. 21. TYPICAL PRESSURE DISTRIBUTION AND WIND DURING A WET WINTER MONTH

The infrabar, 1007 kilobars (29.75 inches), favoring south winds over California

eastern Washington, and eastern Oregon.[1] Similarly, when the Aleutian infrabar extends well southward, individual lows deepen rapidly and move south rapidly. Under such conditions the rain area will extend from the Washington coast to the northern California coast in twelve hours, to the central coast in twenty-four hours, and to the coast south of Point Conception in thirty-six hours.

Relation of individual disturbances to large areas

For the Atlantic coast, Humphreys has advanced the view that unusually mild winters are determined by the persistence of the Bermuda hyperbar, while continued absence of this results in exceptionally low temperatures. Low surface temperature in the vicinity of the Bermudas may depend upon the temperature and strength of the Labrador current, though it is more likely, as we shall see later, that ocean currents are the result rather than the cause of pressure displacement. Humphreys holds that "a persistent strong Labrador current would seem to indicate a subsequent (after a fortnight or longer period) development of a more or less equally persistent Bermuda high, and through it the prevalence during winter of relatively warm weather throughout the eastern United States. On the other hand, a long-continued, weak Labrador current would indicate the subsequent absence of Bermuda highs and the prevalence over the eastern United States of unusually low temperatures."

Effect of Bermuda hyperbar on Atlantic coast

Effect of Labrador current on Bermuda hyperbars

Ward[2] has discussed at some length the cyclonic and anticyclonic control of the weather of the United States. Thus spring and autumn, transition periods, are marked by the struggle between cyclonic and solar controls, and hence by striking convectional phenomena. As summer passes, the sun's rays become more and more oblique and the control of the weather passes gradually, but irregularly, from the sun back again to the cyclone.

Marked convectional phenomena of spring and autumn

[1] For weather maps illustrating movements of California Lows, see E. A. Beals in *Weather Forecasting in the United States*, pp. 327–335; also G. H. Willson, *ibid.*, pp. 335–339.

[2] *Annals of the Assoc. of Am. Geog.*, Vol. IV, pp. 3–54.

21. The effect of ocean currents on atmospheric circulation. Ocean currents belong more properly in a treatise on hydrography, and are appropriately discussed in the publications of nautical institutes, coast surveys, and hydrographic offices connected with the navy departments of various nations. There is, however, a close connection between some of the great currents of the ocean and the great air streams. In water, as in air, there are convectional currents and displacements due to differences in density caused primarily by temperature inequalities; but in water there are also density differences caused by salinity and the origin of the water. Great ocean currents, like great air currents, are deflected by the earth's rotation; and, furthermore, the great sea currents are very often (certainly so far as surface conditions go) driven by the great wind systems. In Figs. 22 and 23, which are essentially Köppen's wind charts, one may get a fair idea of the surface movement of the ocean.

Ocean currents like air currents in many respects

Unlike air currents, ocean currents may be divided into two main classes, warm and cold, and of these the latter are probably the more active in originating and maintaining ocean circulation. Therefore it is proper to begin with the cold Antarctic and Arctic currents. At the south pole there is, as the recent exploring expeditions have shown, a large continent or land mass having an average elevation of 2,200 meters above sea level. Even in the summer months (December, January, and February) the average temperature is below 273°A. In fact, this isotherm almost coincides with the Antarctic Circle (lat. 66° 33′ S.), and this notwithstanding that the earth is then in perihelion. In spite of greater insolation, the farther poleward one goes the lower the temperature falls. The lowest mean annual temperature, that of Framheim headquarters (at sea level, 78° 38′ S., 164° 30′ W.), is 248°A. (−13° F.), the lowest mean on record. Meinardus has called attention to the fact that the east winds south of the trough of low pressure, which were formerly supposed to be of Antarctic origin and have the character of boreal winds, are, on the contrary, moist, warm,

Two main classes of ocean currents

Temperature of Antarctica land mass

snow-producing winds. Except for the humidity and marked precipitation it might naturally be thought that these were foehn winds. In this region, of course, the circulation of the winds in a cyclone is clockwise, or in an opposite direction from that in the northern hemisphere.

The chief factor affecting oceanic circulation is the constant flow of ice from the interior of Antarctica. There would appear to be a constant excess of precipitation over evaporation and consequently a continuous run-off of this snow in the form of glaciers and marginal ice fields. In the summer some of this ice is carried away by the

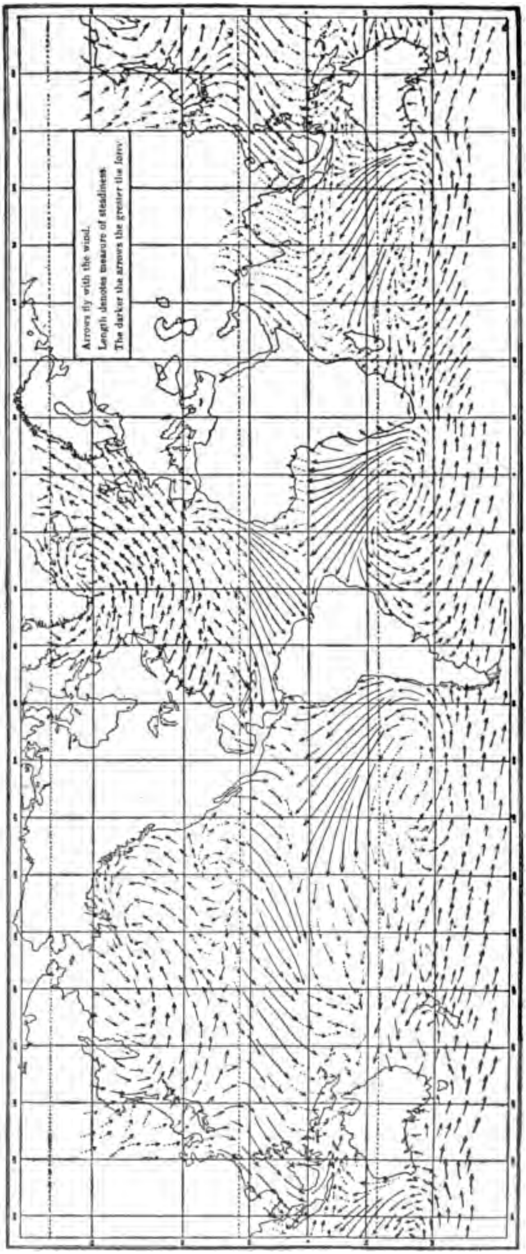


FIG. 22. NORMAL WIND DIRECTIONS AND VELOCITIES FOR JANUARY AND FEBRUARY

After Köppen

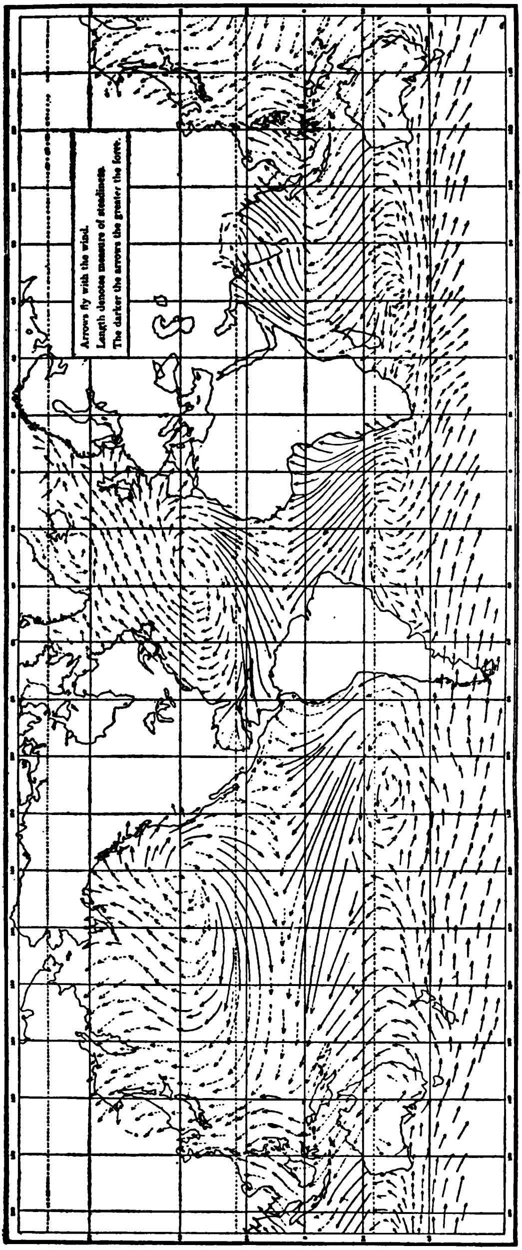


FIG. 23. NORMAL WIND DIRECTIONS AND VELOCITIES FOR JULY AND AUGUST

After Köppen

southern ocean current, which is perhaps the greatest of all sea currents, flowing east around the world in latitude 60° S. Flowing poleward and crossing the main current are the Madagascar, Australian-counter, and Argentine-counter currents, while flowing toward the equator are the Indian, Pacific, Peru, and Georgia currents, and another which joins the Cape Horn current on its way north past the Falkland Islands. Soley (in various reports issued by the Hydrographic Office of the United States Navy) has discussed the basins and flows of these separate currents as well as other currents, and particularly the Gulf

Stream. He states that the great southern ocean current flows eastward around Antarctica sometimes with a velocity of two knots where it narrows to pass through Drake Strait. In November and December immense ice fields separate from the ice barriers, starting generally as large bergs of table shapes and moving with the current; but as the bergs are high out of water the winds from the west accelerate their speed.. In the South Atlantic the direction of the currents is shown very clearly by the movement of the ice, which travels north in the Georgia currents but is cut out very soon where it meets the central counter current.

Direction of South Atlantic currents shown by drifting ice

At the north pole, conditions are different from those at the south pole. Here we have a large, nearly landlocked water surface with an average depth of 2,000 fathoms. Bering Strait and the Greenland Sea are the two openings to the other oceans, and the former is only forty-five miles wide with a depth of twenty-four fathoms. The ocean is frozen nearly all the year. In summer navigation is possible from Point Barrow east for a considerable distance. A current enters through Bering Strait and then divides into three branches, the polar current proper, and the eastern branch and the western branch. On the other side of the Arctic Ocean a great outgoing current is the Greenland current, reinforced by the Spitsbergen current. There is a current flowing into the Arctic Ocean known as the Northeast current, which indeed is an extension of the Gulf Stream. The warm Bering current is in a similar way an extension of the great warm current of the Pacific, namely the Kuro (black) Shio (tide), better known as the Japan current. The Greenland current, which in a way can thus be traced back to the Bering current, joins with the Labrador current, another branch of the Bering, after flowing through Melville Sound and Baffin Bay. The Greenland current follows the east coast of Greenland to Cape Farewell, carrying much floe ice and many large bergs from the Greenland glaciers. At Denmark Strait we have the strange spectacle of two

Currents of the Arctic Ocean

The current through Bering Strait

Bering current

Greenland current

currents flowing in opposite directions within a hundred miles of each other—the Greenland pouring south and the Northeast pouring north. On the east coast of Greenland, temperatures are about 272° A., while on the Iceland coast the temperatures may reach 284° A.

Equatorial currents

The great warm currents are the north and south equatorials, the axes of which are just north of the equator in the Pacific and Atlantic and south of the equator in the Indian Ocean. The Pacific equatorial recurves north of the Philippines in the well-known Japan current, which in its course eastward fans out into what is called the "west wind drift," one branch of this drift, however, going, as we have seen, northward through Bering Strait. The California current is perhaps the return south of the drift.

The Gulf Stream

Between the California current and the coast is a north-flowing eddy current known as the Davidson current. In the Atlantic, the Gulf Stream is the best known of all ocean currents. Temperature observations of it were made as early as 1775 by Franklin. Its importance as a climatic control, especially the widely accepted view that it materially modifies the climate of the British Isles, has been somewhat exaggerated, the true control being the general air movement over a large water surface from west to east and the high specific heat of water. In fact, the current which is marked in the Gulf of Mexico and along the Florida coast fans out and becomes a drift in the middle North Atlantic, a portion of which, as we have seen, becomes the Northeast current, washing Iceland and passing into the Arctic.

Climatic effect of Labrador current

Since the time of Maury the Gulf Stream has been much studied by navigators. Lieutenant Soley of the United States Navy (in charge of the Hydrographic Office at New Orleans) has contributed much to recent knowledge of this stream, its variations in intensity, and its direction. Probably the Labrador current has a more positive and direct climatic effect upon the surface temperature of the eastern part of the North Atlantic than is generally supposed. Changes in the

strength of the trades in the Atlantic and the relation of North Atlantic surface temperatures to the weather of the British Isles have been studied by Commander Hepworth, R. N. R., and the results published by the Meteorological Office of Great Britain.[1] Particularly valuable are the tables of the North Atlantic mean temperature for the surface for the year, the monthly isotherms, the thermoisopleths for surface temperature from the Florida straits to Valencia, and the charts of ice frequency and departures of pressure and temperature for the five-year period beginning 1908 and ending 1912.

Ice conditions in Labrador current

Since 1913 special cruises have been made by officers of the U. S. Revenue Cutter Service with a view to studying ice conditions near the Grand Bank and on the Labrador Coast. The main purpose was to secure observations across the Labrador current. During the early summer of 1914 ice conditions on the Labrador coast were the worst for years and vessels specially constructed to withstand heavy ice were unable to make any headway. The farthest point north reached by the *Seneca* was 52° 45′, longitude 53° 15′. The largest iceberg seen on this cruise was nearly 200 meters long, 180 meters wide, and 15 meters out of the water, one great block of ice weighing about five million tons. Scattered about were other bergs, the average being about one in every five miles. South of latitude 50° there was a berg to be seen about every ten miles on the average. It was calculated that there were about two thousand bergs between Hamilton Inlet and the Grand Bank.

In future surveys, it may be feasible to employ hydroplanes, permitting the mother ship to remain at a safe distance, while the extent of the ice field is being determined. In foggy weather also, the plane can rise above the fog.

The currents on the Grand Bank appear to be tidal and largely influenced by the wind. The tide begins to flood from the southward and ends from the northwest; it begins to ebb from the northward and ends from the southeast, going round in almost a complete circle.

[1] *Geophysical Memoirs*, 1, 1912, and 10, 1914.

In the opinion of Captain C. E. Johnston of the U. S. Revenue Cutter Service, the results of the two cruises made by him in these waters show that, during June, the Labrador current dwindles to nothing and the northern branch of the Gulf Stream spreads out, fan-shaped, losing depth and temperature and becoming a surface drift, which gradually carries the ice northward of the Straits of Belle Isle; there part of the ice melts and the rest remains until the next spring, when it comes south again as the Labrador current gains strength.

Labrador current in June

CHAPTER VIII

The Minor Circulations

22. Cyclones and anticyclones. The word "cyclone" was introduced by Piddington in the first edition of the *Sailor's Horn-book* published in Calcutta in 1848. It is from a Greek word signifying, among other things, the coil of a snake and was used as neither affirming that the air moved in a true circle, nor that the circuit was complete, but still expressing the tendency to circular motion in the typhoons of the East Indian waters.

Franklin's observations

Franklin had a clear idea of the progressive movement of storms, as is shown in his letter of July 16, 1747. In another letter, dated February 13, 1750, referring to an eclipse of the moon on October 21, 1743, Franklin describes a severe northeast storm which prevented his observation of the eclipse and his surprise in finding that observations had been made in Boston. He says, "I wrote to my brother about it and he informed me that the eclipse was over an hour before the storm began."

Capper in 1801 described in his *Winds and Monsoons* certain hurricanes as revolving masses of air; and Horsburgh in 1805 described the typhoons of the China Sea as rotary storms. The first American storm to be charted was that of September 23, 1815, by Professor Farrar of Harvard. In 1831 Redfield published the first of a series of papers demonstrating what was then called the Law of Storms. Since then, much has been written on the mechanics of cyclones; and various theories have been advanced as to their origin, maintenance, and motion. Dove, Piddington, Reid, Redfield, Espy, Ferrel, Hann, Davis, Bigelow, Shaw, and others have contributed to the subject; but with the introduction of new facts relating to the upper air, the various explanations have been proved inadequate. The earlier explanations assumed a movement of the air in circles, and an active cause in the condensation of vapor with uprising currents; but it now is definitely known that these are of secondary importance. By "cyclone" we

mean the familiar storm of temperate latitudes, a large aërial whirl or eddy, with a diameter of several hundred kilometers.

Direction of cyclonic whirl

In the northern hemisphere the rotation is in a direction contrary to the direction of the hands of a watch, while in the southern hemisphere the movement is in the same direction as the hands of a watch. In the center of the whirl the pressure is lower than at the periphery, and so this is generally spoken of as the storm center, or depression center, although it is by no means certain that this lowest pressure, as generally obtained and charted, is the true center of gyration. Storm tracks plotted upon the assumption that the line joining successive lowest pressures indicates the movement of the center of air motion, are misleading, particularly so when over-large corrections for surface temperature have been used in reducing pressure readings.

The center of gyration

From studies of storms in the United States, Bigelow holds that there are no true local warm-centered and cold-centered cyclones or anticyclones; and that all theoretical discussions founded on such a basis are misdirected. Observations demonstrate that in the lower atmosphere the actual mechanism consists of rather deep, warm and cold counter-currents of air underrunning the prevailing eastward drift. The centers of gyration are uniformly in the region where these counterflowing currents meet; that is, on the edges rather than in the midst of the warm and the cold regions. The front half of the cyclone is relatively warm and the other half cold; while the rear half of the anticyclone is warm and the front half cold.

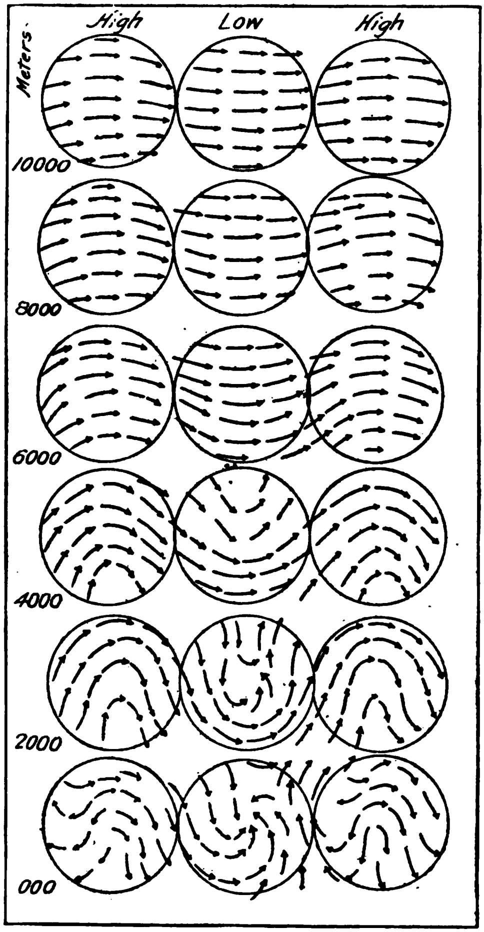
Mechanism of the lower atmosphere

Stratification and interpenetration of currents of different temperatures may be the true source of the energy of storms. The heat energy derived from the condensation of aqueous vapor, and the energy produced by purely dynamic eddies, are entirely secondary in importance to the energy obtained from the counterflow and underflow of warm southerly currents against the cold northerly currents and beneath the eastward-flowing drift. There appears to be no difference in the structure of European and American cyclones and anticyclones. Instead of vertical

Source of the energy of storms

convection and condensation being the prime movers, it rather is a matter of horizontal convection and interchange of heat on the same general level. There is no evidence, according to Bigelow, of the superposition of cold-center cyclones upon warm-center cyclones as expounded by some earlier writers, nor are there any purely dynamic vortices in a rapid stream as supposed by Hann, nor are there cyclonic vortices caused by atmospheric islands of high pressure obstructing a rapidly flowing eastward drift.

The cyclonic circulation constitutes an effort to bring back to equilibrium the energy difference represented in cold and in warm areas; and this is done by setting up an extensive series of internal vortices varying in size from the large storm areas down through tornadoes or secondaries to the minute whirls one sees in the desert as dust whirls or on the sea as waterspouts.



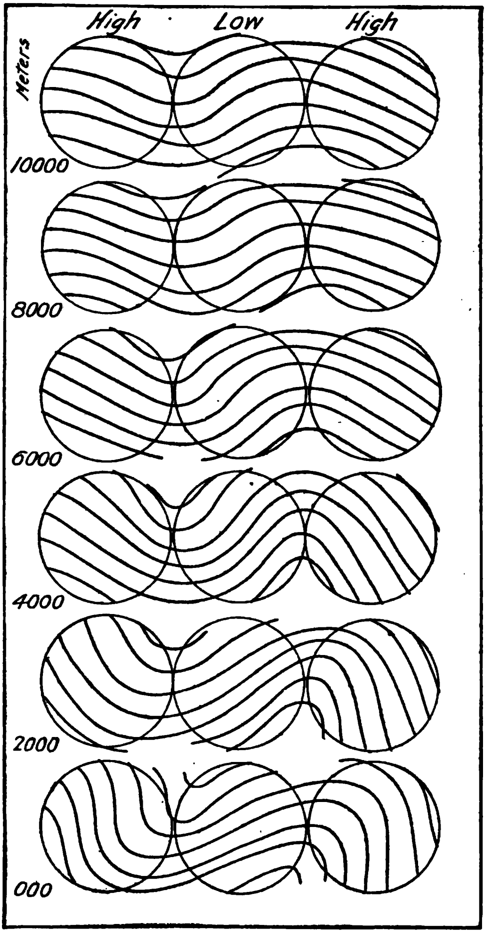
After Bigelow

FIG. 24. TYPICAL DISTRIBUTION OF THE WIND IN CYCLONES AND ANTICYCLONES

In ordinary cyclones the vortices are not perfect, and it is only rarely and in highly developed, localized storms, such as the tornado and the waterspout, that continued vortex motion is attained. Discussion of these types will be given

later. For diagrams of velocity and temperature in cyclones see Figs. 24, 25, and 26.

W. H. Dines, studying the high-level records, found that differences of pressure at the earth's surface were of the same order of magnitude as those at a height of 9 kilometers; and he drew the inference that the surface distribution was mainly controlled by the conditions at the higher level. Shaw has carried this view further and demonstrated that the intermediate layers contribute little to the distribution at the surface and that the stratosphere, although made up of only one quarter of the atmosphere, is the dominant factor; while the troposphere, though it includes three quarters of the atmosphere, has comparatively little influence. This difference in the influence of the two great layers is, Shaw thinks, attributable simply to the characteristic difference of temperature between regions of high and of low pressure. He has deduced two equations, one giving the increase of pressure difference in kilobars per meter of height, and the other for the gradient velocity.

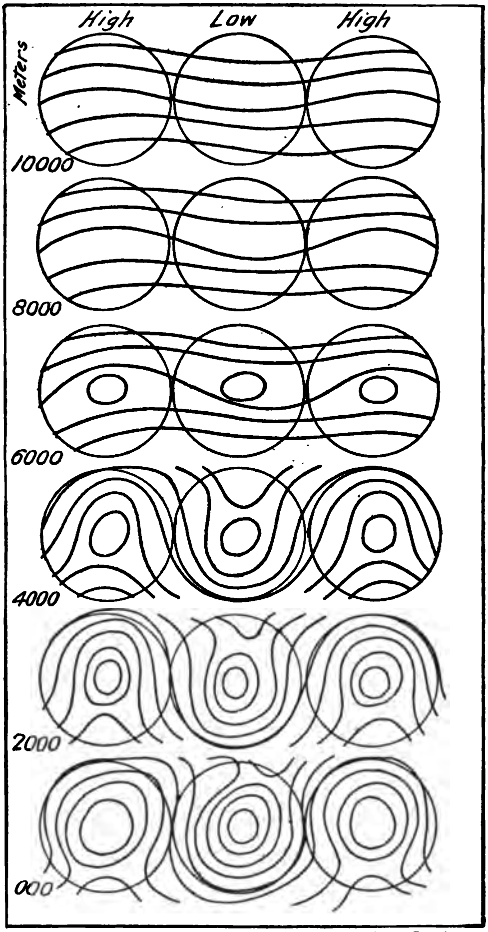


After Bigelow

FIG. 25. TYPICAL DISTRIBUTION OF THE TEMPERATURE IN CYCLONES AND ANTICYCLONES

Among other things, he has proved that if certain assumptions are granted the distribution of pressure and temperature can be computed from measurements of wind velocity at certain heights, such as are obtained from the observations of a pilot balloon.

A number of cases are discussed where the slope of the temperature is in different directions, also cases of inversion and comparison made with types of wind structure as given by Cave. What is called the operative distribution of pressure, at a height of about 9 kilometers or the level at which the wind velocity is greatest, is used in connection with "Egnell's" or Clayton's law,[1] that is, the law of equal mass transport at all heights, or wind velocity increasing as density decreases, to show that up to the highest cloud level the amount of air carried by each kilometer layer is the same. It would appear, then, from the pilot-balloon soundings, that the maintenance of the pressure gradient and consequent



After Bigelow

FIG. 26. TYPICAL DISTRIBUTION OF THE PRESSURE IN CYCLONES AND ANTICYCLONES

[1] This law was first demonstrated in connection with the cloud observations at Blue Hill Observatory by Clayton.

increase of velocity aloft are really due to the influence of the thermal structure of the underlying atmosphere upon the pressure distribution transmitted from above. The most general characteristic of that structure is a slope of temperature toward the north. A general statement that the velocity always increases aloft would certainly be subject to many exceptions, but it seems correct to say that the usual thermal structure of the atmosphere is such that *the westerly component of the air current* is greatest in the operative pressure layer and gradually diminishes from there to the surface.

General characteristic of the thermal structure

Shaw[1] gives five laws of atmospheric motion which seem to him to be fundamental. These are:

1. *The law of the relation of motion to pressure.* In the upper layers of the atmosphere the steady horizontal motion of the air at any level is along the horizontal section of the isobaric surfaces at that level, and the velocity is inversely proportional to the separation of the isobaric lines in the level of the section.

This law is not experimentally demonstrated and it is quite possible it may be of limited application. It presupposes a condition of steady motion. Shaw points out that allowance must be made for "curvature of path" which will vary with latitude. For half of the globe north of 30° N. and south of 30° S. it is generally negligible, but near the equator it becomes the paramount consideration in the question of the persistence of distribution.

Five laws of atmospheric motion

> "Thus, rotary systems, small or large, are the only possible isobars for a synchronous chart of an equatorial region, if one were drawn. Long sweeps of parallel isobars would be impossible there."

2. *The law of the computation of pressure and of the application of the laws of gases.*

This is simply an amplification of the well-known equation of elasticity.

3. *The law of convection.* Convection in the atmosphere is the descent of colder air in contiguity with air relatively warmer.

[1] *Principia Atmospherica.*

4. *The law of the limit of convection.* Convection in the atmosphere is limited to that portion of it called the troposphere, in which there exists a sensible fall of temperature with height. The upper layer of the atmosphere, in which there is no sensible fall of temperature with height and therefore no convection, is called the stratosphere.
5. *The law of saturation.* The amount of water vapor contained in a given volume of space cannot exceed a certain limit which depends upon the temperature.

As a postulate, based on Dines' statistical studies, it may be stated that in the stratosphere from 11 kilometers upward, it is colder in the high than in the low at the same level; and in the troposphere, from 9 kilometers down to 1 kilometer, it is warmer in the high pressure than in the low at the same level. The average horizontal circulation in the northern hemisphere in January between 4 kilometers and 8 kilometers consists of a figure-of-eight orbit from west to east along isobars around the pole, with lobes over the continents and bights over the oceans. The average circulation at the surface is the resultant of the circulation at 4 kilometers combined with a circulation in the opposite direction of similar shape due to the distribution of temperature near the surface.

The strophic balance

Shaw in a recent lecture[1] on "Illusions of the Upper Air," using what he has called the principle of strophic balance, shows that this has the great advantage of giving a definite relation between wind velocity, pressure, and temperature. Using the familiar symbols

- p for pressure
- θ for temperature
- ρ for density
- l for horizontal distance
- h for vertical height
- s for horizontal pressure gradient
- q for horizontal temperature gradient
- v for velocity of wind, positive when pressure is high on the right of the path
- E for the radius of the earth
- g for normal acceleration of gravity
- r for the angular radius of a small circle on the earth's surface which indicates the path of air in a cyclone

[1] Lecture before Royal Institution, March 10, 1916; *Nature*, April 27, 1916, p. 191, and May 4, 1916, p. 210.

ϕ for latitude
ω for the angular velocity of the earth's rotation

he obtains for the fundamental relation between the velocity of the wind at any level and the pressure gradient there

$$\frac{dp}{dl} = s = 2\,\omega v \rho \sin\phi \pm \frac{v^2}{E}\rho \cot. r.$$

The two terms which make up the right-hand side of this equation are of different importance in different places and circumstances. If the air is moving in a great circle, r is 90°, and cot. r is zero; so that the first term alone remains. At the equator the latitude is zero and the second term alone remains. Away from the equatorial region the second term is relatively unimportant unless the velocity is great. In temperate and polar latitudes the path of the air differs little from a great circle except in rare cases near the center of deep depressions; consequently the first term may be regarded as the dominant term in these regions.

Geostrophic wind

Cyclostrophic wind

Shaw calls the wind computed from the first term the *geostrophic* wind or practically the actual wind of temperate and polar region. The wind computed according to the second term is called the *cyclostrophic* wind or the actual wind in equatorial regions, or the wind of tropical hurricanes.

By simple manipulation of the fundamental formulae including the characteristic equation Shaw obtains the following:

For change of pressure gradient with height $\left\{ \dfrac{ds}{dh} = g\rho\left(\dfrac{q}{\theta} - \dfrac{s}{p}\right) \right.$

For change of wind velocity with height: geostrophic winds $\left\{ \dfrac{dv}{dh} = \dfrac{vd\theta}{\theta dh} + \dfrac{g}{2\,\omega\sin\phi}\cdot\dfrac{q}{\theta} \right.$

cyclostrophic winds $\left\{ \dfrac{dv^2}{dh} = \dfrac{v^2 d\theta}{\theta dh} + \dfrac{gE}{\text{cot. } r}\cdot\dfrac{q}{\theta} \right.$

From these equations he deduces what may be called the law of equivalence of pressure distribution and wind which in his opinion also serves to explain the following facts established by observation:

Light winds in the central region of an anticyclone: It follows from the fundamental equation when the negative

sign is taken, as it must be for an anticyclone, that the values of v will be given by the roots of a quadratic equation, which will be impossible if v is greater than $\frac{E\omega \sin\phi}{\cot. r}$. This for a radius of 113 kilometers (70 miles) allows a velocity of only about 4 meters per second. This, Shaw holds, furnishes a crucial test, for if an anticyclone is a region of descending and outward flowing air, the velocity should diminish as the air spreads outward. In an anticyclone the reverse condition occurs. This, however, is open to the criticism that in practice advancing cyclones may mask the true circulation and velocity.

The small influence of the troposphere, and therefore the dominance of the stratosphere in the distribution of surface pressure, follow in Shaw's opinion when numerical values are given in the equation for change of pressure gradient with height or

$$\frac{ds}{dh} = g\rho\left(\frac{q}{\theta} - \frac{s}{p}\right).$$

The right-hand side of the equation consists of two terms which are of opposite sign and numerically nearly equal in the middle regions of the troposphere. Their combined effect for the whole range is therefore relatively small and the change of pressure produced in the troposphere is unimportant. The distribution of the stratosphere is dominant throughout the troposphere.

Again, the apparently capricious variations of wind and temperature with height shown by pilot balloons and *ballons-sondes* may be connected numerically by the equation for the change of pressure gradient with height and the equation for horizontal pressure gradient. Shaw gives an example where the rapid transition from a southerly wind at 1,100 meters through a calm to a northerly wind at 1,500 meters on October 16, 1913, indicated a temperature gradient of 7° per hundred kilometers toward the east. But this condition was in satisfactory accord with the meteorological circumstances at the time. The same combination of equations enables us to specify the *conditions under which "Egnell's law,"* that wind velocity at different heights is inversely

proportional to the density at those heights, may be expected to be verified, and the conditions prescribed are essentially reasonable.

The rapid falling off of wind in the stratosphere noted in observations with pilot balloons follows from the equation for change of wind velocity with height.

The same equation applied to the troposphere, *assuming normal values for temperature*, gives correctly the rate of change of velocity with height. The permanence of vertical motion about a vertical axis in the atmosphere follows from the equation for cyclostrophic winds. With a wind velocity of 20 meters per second and a horizontal temperature gradient of 5° per hundred kilometers, the extension will be 1.4 kilometers upward; so that the vortex will be covered by a cap in which the velocity gradually falls off to zero within a very limited height.

For the extension downward the calculation is more complicated; but the computed change of velocity is very small so that the vortex must be regarded as reaching the ground; and it would appear that a vortex extending throughout the troposphere terminating with a cap in the stratosphere is a possible reality.

Thus the hypothesis of an atmosphere in which the wind velocity is everywhere adjusted to balance the pressure distribution, enables us to explain many of the ascertained facts that have been disclosed by the investigation of the upper air, and strongly supports the idea that the pressure distribution at the surface is controlled by the stratosphere and only modified locally by convection.

CHAPTER IX

Forecasting Storms

23. Types of storms. It is the practice of the Forecast Division of the Weather Bureau to classify storms after the region where first charted. Thus, one of the most frequent types is known as the "Alberta," because it is definitely charted in that territory. The system has its defects, and with each enlargement of the area of observation some modification of the place of origin becomes apparent. In a recent publication by Bowie and Weightman[1] the types are given as Alberta, North Pacific, South Pacific, Northern Rocky Mountain, Colorado, Texas, East Gulf, South Atlantic, Central, and West Indian. Charts showing the normal twenty-four-hour movement for five-degree squares have been prepared to take the place of the earlier charts of paths of greatest frequency. Tables covering a period of twenty years are also available for the average velocity. Thus it is seen that storms of continental types move more rapidly in winter than in summer.

Type of storm named for place of origin

	Kilometers in 24 Hours	Miles
January	1,199	745
February	1,110	690
March	1,080	673
April	871	542
May	792	492
June	772	480
July	839	521
August	788	489
September	883	549
October	919	571
November	1,040	646
December	1,156	718
Average for year	954	593

[1] *Monthly Weather Review*, Suppl. 1, July, 1914; also Suppl. 4, Jan., 1917.

In determining a possible deviation from a normal course, account is taken of unequal pressure distribution in the regions adjacent to the storm center, also the location of maximum twelve-hour pressure fall and the trend of the isotherms. A number of empirical rules based upon the intensity and movement of the twelve-hour pressure change are given by Bowie for the movement of lows.

Empirical rules for movement of lows

The most important rules[1] for the guidance of the forecaster in determining the course of a hurricane are:

"A hurricane does not move directly toward a region of high pressure when such an area is not moving perceptibly, but follows behind it. If the high moves east or northeast off to sea at a normal rate, the hurricane moves north or northeast. If the high hangs persistently over the coast, the hurricane is deflected far to the west before it can recurve.

West Indian hurricanes

"If rain falls freely before the hurricane comes to land, the disturbance may decrease in intensity; but if heavy rain begins after the storm passes inland, the storm will probably continue.

"When a West Indian hurricane is moving westward in the longitude of eastern Cuba and is north of that island, it will recurve east of the South Atlantic coast, when an area of high pressure covers the northwestern states. If the hurricane is moving westward over Cuba or the western Caribbean Sea when a low area occupies the northwest, and the pressure is high in the eastern states, the storm will probably move to the Gulf of Mexico and reach the Gulf coast after recurving.

"For storms over the Great Lakes, it appears that depressions frequently remain stationary or move slowly, accompanied with much precipitation, when the pressure is high to the north and northeast. Again, the movement is slow when the air from an extensive high-pressure area drains southeast from the Missouri Valley.

Storms over the Great Lakes

"Other storms that increase with intensity appear to depend

[1] In *Weather Forecasting in the United States* many general statements are made by the various forecasters regarding the movements of HIGHS and LOWS.

on marked horizontal temperature gradients. A rapid temperature rise in front of a storm implies an increase in intensity, especially if the temperature is falling rapidly over the northwest. Sharp temperature rises in the eastern quadrants of a storm are a sure indication that the storm will move northeastward and increase in intensity."

March storms

On the Atlantic coast there are certain types of disturbance which, especially in March,[1] have provoked widespread comment owing to the failure of the Washington forecasters rightly to anticipate weather conditions of the succeeding twenty-four hours. Noteworthy instances were those at the time of the presidential inaugurations in 1897 and 1909.

Auxiliary maps

For forty-five years the forecasters at Washington and for shorter periods at other forecast centers such as San Francisco, Chicago, New Orleans, Portland, and Denver have depended mainly in making their forecasts upon certain auxiliary maps of pressure and temperature changes. Attempts have been made to use cloud change and humidity change maps, but for reasons hardly satisfactory, it would seem, no continuous use of these latter charts has been made. The pressure and temperature auxiliary maps show the 24- and 12-hour changes. The *barometric tendency*, as defined by the International Committee in 1913, is the change in the three hours preceding the observation. This is not used in the United States, but whenever the pressure has risen or fallen 1.02 millimeters within two hours preceding the observation, the change is reported, though not the character of the change, such as steady, unsteady, etc.

The barometric tendency

The pressure chart gives the area and intensity of the nonperiodic pressure changes. Henry, compiling the fluctuations in pressure at certain stations for a period of ten years, found that the frequency was nearly the same for all parts of the country except that the changes are more rapid at northern stations. The average annual number of such pressure movements is 88 and the average time interval 4.2 days.

[1] The snow and wind storm known as the Great Blizzard of New York occurred March 12–14, 1888.

Ekholm in 1911 suggested the terms *allobar* for the area of pressure change; *anallobar* for an area over which the pressure has risen; and *katallobar* for an area over which the pressure has fallen within the given time. The region of maximum change may be regarded as the center. The names are cumbersome and the conditions might well be described simply as "rises" and "falls." Henry[1] has described the basis of forecasting by synoptic maps, and given at some length the relation of the pressure change areas and the movements of highs and lows. Shaw[2] has discussed certain relations between the isallobars and the winds.

Allobars

Storms over the Lake region sometimes develop secondaries off the Virginia or New Jersey coasts; and these pass apparently slowly northward, causing heavy snows and high winds in the Middle Atlantic and New England states.

Secondaries

Of all secondaries, tornadoes are most destructive and most frequent. They are associated with storms of increasing energy, moving to the left of normal paths when the trough of low pressure extends well southward. (For a good illustration see the weather map of April 29, 1909.) Again, when the southern portion of the trough swings eastward faster than the northern portion, there is likelihood of the development of a secondary storm south or southeast of the northern center. (See map of November 8, 1913.)

There is a tendency for secondaries to form to the leeward of the Appalachian Mountains, following the passage eastward of moderate disturbances from the northwest. A pressure rise coming from the Lake region seems to play an important part. If this moves south of the low, secondaries do not develop.

For apparent paths of lows and highs in the United States as given by Van Cleef see Figs. 27 and 28.

24. Tornadoes. These are secondary and localized vortices of great energy, forming as a rule in the southern quadrants of a larger cyclonic storm. In the southern hemisphere this relation is reversed. The characteristic feature is a

[1] *Weather Forecasting in the United States*, p. 69.

[2] *Forecasting Weather*, pp. 337–341.

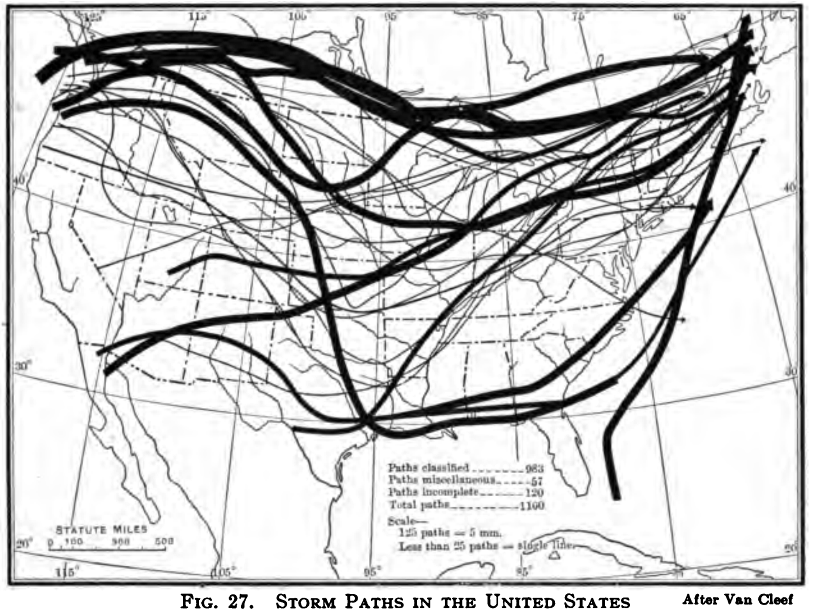


FIG. 27. STORM PATHS IN THE UNITED STATES After Van Cleef

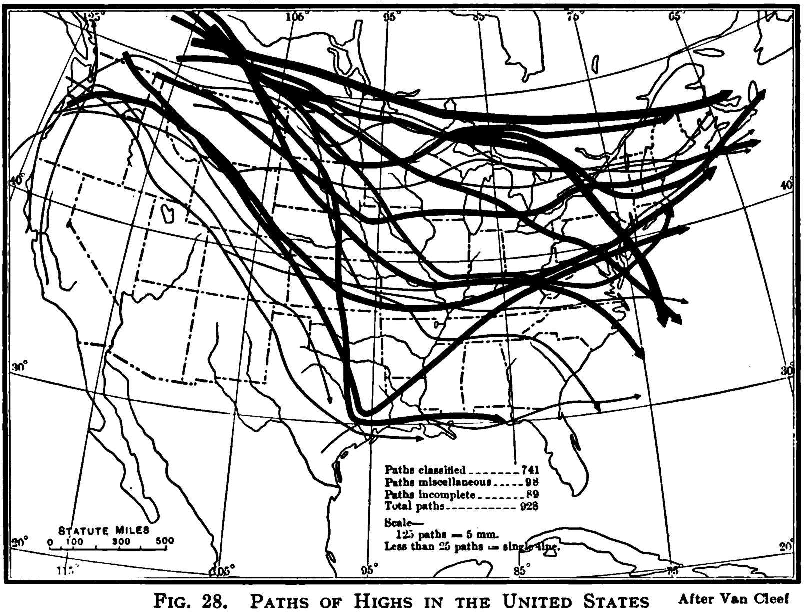


FIG. 28. PATHS OF HIGHS IN THE UNITED STATES After Van Cleef

funnel-shaped cloud, although the term is widely used for any marked wind storm of great violence. The word was originally used for certain storms off the African coast in which there was a quick change of wind direction. The word is used in the press as equivalent to "twister," and often wrongly used as the equivalent of cyclone. Tornadoes are best defined as extremely violent vortices with funnel-shaped clouds, the diameters of which vary from 100 to 500 meters, while the larger storms (cyclones) of which they are a part have diameters a hundred times greater. Owing to centrifugal force, the pressure is very low within a limited area; and as ordinarily stated, a partial vacuum is developed. The destructive effects therefore are due both to the extremely high velocity of the wind, which may reach 100 meters per second, and to the rapid change in pressure. The vortex moves east or northeast at a rate of 10 or 15 meters per second and may preserve its identity for an hour or longer, and it is not unusual for several tornadoes to form within a few kilometers of each other and move in parallel paths. These storms are very destructive, and phenomena that seem almost beyond belief occur during their passage over a given spot. Thus chickens have been stripped of their feathers, large animals have been lifted and carried some distance, straws, twigs, and other small articles have been driven into boards, owing to their high velocity, notwithstanding their small mass. The walls of buildings fall outward as in an explosion, and trees are uprooted, falling as the whirling air leaves them. Thus on the north side of the path of the center, trees are thrown to the southeast, while on the south side they are thrown to the northeast. Tornadoes occur under the same general conditions as thunderstorms, along the line of conflict between warm, moist south winds and cold north winds. They are frequent in the spring.[1] Grinnell, Iowa, was devastated by a tornado on June 17, 1882, and Rochester, Minn., on August 21, 1883. Tornadoes of historic interest occurred at

Characteristic feature

Cause of their destructive power

Effects of tornadic activity

[1] For good examples of tornado conditions, see the weather maps of Feb. 19, 1884, and Mar. 23, 1913.

Lawrence, Mass., July 26, 1890 (Fig. 29); at St. Louis, May 27, 1896; and at Omaha and various points in Nebraska and adjoining states on March 23, 1913. Finley has discussed statistically this type of storm for the United States. The storm as an example of vortex motion has been studied by Bigelow, who has computed the radial and tangential velocities at various points for the St. Louis tornado of May 27, 1896.[1] Briefly, the vortices extended upward about 1,200 meters and had a diameter of

Vortices of certain tornadoes

FIG. 29. TORNADO AT LAWRENCE, MASS.

The center of the tornado passed along the side of the street nearest the overturned house. The house just beyond it was drawn forward six or eight feet to the edge of the sidewalk shown by the small tree.

1,900 meters near the ground. The velocity of the air near the center of the vortex at the ground exceeded 250 meters per second.

25. Waterspouts. These are small secondary storms occurring generally at sea, although there are instances of their occurrence inland over water courses. Thus on July 16, 1904, a well-defined spout was observed by many people over the Hudson River near Tarrytown. This is described by M. L. Bacon,[2] who succeeded in getting six photographs

[1] *Monthly Weather Review*, Aug., 1908.

[2] *Ibid.*, June, 1906.

of the spout. Apparently the cloud did not descend in a funnel shape to meet the rising column of water; but the column of water rose, clear cut, and met the cloud, forming a vertical column at 4:25 P.M. At 4:35 P.M. the spout had disappeared.

Conditions preceding waterspouts

A waterspout which occurred off Cottage City in Vineyard Sound on August 19, 1896, has been studied in detail by Bigelow,[1] who came to the conclusion that it was caused by a sheet of cold air overrunning the low, warm, quiet strata, about midday; while the cold air followed at the surface a few hours later. In such facts we have the conditions required to produce marked vertical convection,—sudden cooling of the upper strata and an abnormal stratification.

The following description of a triple waterspout is given by F. A. D. Cox, lieutenant in the Royal Navy:[2]

"On November 10, 1912, about 10 A.M. . . . bright sunshine, when gradually the sky became overcast and there

Photograph by Chamberlain

FIG. 30. THE GREAT WATERSPOUT EIGHT MILES OFF COTTAGE CITY, MASS., AUGUST 19, 1896

was a dense rain cloud above as shown in the photograph. Below the bank of clouds to sea level it was quite clear and

[1] *Monthly Weather Review*, July, 1906.
[2] *Symons's Meteor. Magazine*, April, 1913.

F. A. D. Cox in Symons's Meteor. Magazine

FIG. 31. WATERSPOUTS AT CHATHAM ISLANDS

bright and toward the east heavy rain was evidently falling. A little after 11, my attention was called to a peculiar funnel-shaped cloud which was beginning to appear on the lower edge of the cloud bank. I saw at once that something out of the common was beginning and said: 'Well, I have never seen a waterspout, but it looks to me as if this was one forming.' We then carefully watched and it was soon plainly evident that a waterspout was taking place. We marked how the funnel-shaped excrescence from the cloud bank gradually extended downward to the sea, and from below we could observe another funnel rising which soon joined the one above; the whole appearance had that of a spiral tube evidently formed by a rotary motion; the water on the sea end of the spout was in a perfect foam. The spout first formed was to the right hand on the eastward side next to where the heavy rain was evidently falling. Almost immediately after the formation of the first spout another began to make its appearance. It was much thicker than the first, and as in that case the sea below the cloud was violently agitated, and even from where we stood, which must have been seven or eight miles distant, the form, like columns, was plainly seen. This was by far the largest of the spouts and continued for nearly half an hour. Toward the end, before the spout began to subside, the sea had almost the appearance of a geyser, so violently was it agitated. I think the large one must have been nearer to us than the first, for as the first began to dissolve, it gradually drifted toward the big one, and soon the remains appeared as a sort of appendix

Triple waterspout

hanging from the cloud above. The spouts disappeared slowly, and the whole phenomenon occupied about three quarters of an hour."

Thunderstorms are also secondary storms; but owing to the marked electrical phenomena accompanying them, the discussion is more properly placed under the section on atmospheric electricity.

CHAPTER X

THE WINDS

26. Wind systems. Edmund Halley, the astronomer, in 1698 received from King William III command of a sailing ship, with directions to study variations of the compass. He made two memorable voyages and practically covered the Atlantic from 50° N. to 50° S. Besides the magnetic work much meteorological work was done, and our first knowledge of the general wind system of the Atlantic comes as a result of these voyages.[1]

First knowledge of wind system of the Atlantic

In 1856 another explorer, Captain Charles Wilkes of the United States Navy, read a paper before the American Association for the Advancement of Science which was later published. It is in this work that we find included one of the earliest maps of the winds of the world. In 1875, through the joint agency of the Smithsonian Institution and Professor J. H. Coffin of Lafayette College, there was published a large volume on *The Winds of the Globe*. Several world maps of wind movement are given; and not only the annual direction but the directions for summer and winter months are charted. Köppen of Hamburg has given us the latest of these charts.

Early maps and charts of winds

It is apparent that there are several mighty streams of air flowing around the world in certain latitudes. Some blow steadily and are more or less permanent, like the trades, the anti-trades, and the prevailing westerlies. Some are seasonal in character, like the monsoons. There are also well-marked minor circulations, known as sea breezes and valley winds; and finally, there are the individual, localized winds accompanying the various storm types.

Various types of winds

The permanent, or planetary, winds are controlled by the planetary pressure distribution; that is, they depend upon the

[1] See section 18.

general difference of temperature between equatorial and polar regions, and more especially upon the position and strength of the great planetary pressure belts. The seasonal winds can be correlated with movements of the hyperbars or infrabars (the so-called centers of action). The local winds can be traced to temporary disturbances of pressure.

Agencies controlling the winds

The winds have been classified by Dove, Davis, and others as planetary, terrestrial, continental, land and sea breeze, mountain and valley breeze, cyclonic, and certain accidental winds due to volcanic eruptions.

27. **Trade winds.** These are the great northeast and southeast wind systems. The name is derived from the old English, to blow trade, meaning in one direction. On the pilot charts issued by the Hydrographic Office founded upon the researches made and the data collected by Lieutenant M. F. Maury, there is published the average condition of wind and weather for the given period. Thus if we look on the chart of the North Pacific Ocean for May, we find that the northeast trades, force 4 to 5 (5 to 8 meters per second), extend to within about 5° of the American coast between the 25th and 15th parallels. They average twenty-four days in May over the Hawaiian Islands. On the other hand, the southeast trades, force 3 to 4 (3 to 5 m. p. s.), extend 1° to 5° north of the equator, and are farthest north between longitudes 150° W. and 110° W.

Spring trades of the Pacific

Over the Atlantic during May we find that the northeast trades extend northward slightly beyond the Canary Islands, but west of the 30th meridian the northern limit of these winds is nearly along the 25th parallel. The southern limit is close to the equator on the American side, but rises to latitude 12° N. at longitude 20° W. The force of the northeast trades is 4 to 5, increasing toward the south. Their direction is northerly off the African coast, but is northeast between the 20th and 30th meridians. Farther westward the direction is more easterly, and north of the Lesser Antilles it is southeasterly, showing the anticyclonic circulation around the Azores hyperbar. The winds

Spring trades of the Atlantic

are generally east to northeast in the Caribbean Sea and east to southeast in the Gulf of Mexico. The southeast trades, force 3 to 4, extend from 1° to 30° above the equator between the 8th and 42d meridians.

Spring trades of the Indian Ocean

Over the Indian Ocean during May the southeast trades prevail over the area between the equator and latitude 30° S. Over the extreme northern and southern portions of this area the trades are broken by variable winds, and calms are frequent between the equator and 10° S. The trades are steadiest between latitudes 10° S. and 25° S. Along the African coast, near the equator, they follow the contour of the land and merge into the southwest monsoon.

Winter trades of the Pacific

If we follow the trades during winter months we shall find that over the North Pacific the northeast trades reach their most northern limit in the eastern part of the ocean at the 29th parallel, slightly southeast of the central area of the California high, and are strongest and steadiest south of this region. Between longitudes 145° W. and 155° E. their northern limit is close to the 25th parallel. They extend eastward to within 5° to 8° of the American coast and westward to Asiatic waters, where they merge into the northeast monsoon. They extend as far south as the equator west of longitude 170° E. East of this longitude their southern limit gradually rises to the 10th parallel at longitude 125° W. The southeast trades extend north of the equator between longitudes 85° W. and 180° W. They reach their most northern limit, the 6th parallel, between longitudes 115° W. and 125° W. In the North Atlantic the northeast trades prevail between the 5th and 25th parallels. Near Brazil they extend as far south as the equator; and near the African coast as far north as latitude 32° N. These winds are the typical northeast trades over the eastern part of the ocean and in the Caribbean Sea. In the central part of the ocean they become east-northeasterly. Southeast trade winds extend north of the equator over the central part of the ocean to the 4th parallel.

Winter trades of the Atlantic

In the South Atlantic the southeast trades prevail from the

area of high pressure to latitude 5° S. on the eastern part of the ocean, and from latitude 15° S. to the equator on the western. Over the greater part of this area they are well developed, blowing from the southeast from 50 to 60 per cent of the time, with a small percentage of calms and no gales, the average force being about 4. South of the area of high pressure, "the brave west winds" prevail. They have increased slightly in intensity since the spring. The winds around the high show their anticyclonic movements very plainly, while those within the area are variable in direction and force. Over the Indian Ocean in winter the southeast trades occur between 10° S. and 30° S. east of the 50th meridian. Their average force is 3 to 4. West of Madagascar the winds are mostly northeasterly and southeasterly, while south of Madagascar they are easterly.

Winter trades of the Indian Ocean

In discussing the planetary circulation it was assumed that in the upper levels there must be an overflow of air from the equator to the poles. Recent soundings, however, do not confirm this view. And this variability in flow is shown in marked degree above the trades. The trades themselves are comparatively shallow streams, not extending above the 5-kilometer level. Above these the air movement is from west to east; and these winds are called the anti-trades, somewhat unfortunately since this term is applied to the winds farther north or south, better described as the prevailing westerlies. We shall use the term "counter-trades" for the winds above the trades. The counter-trades, then, are above the trades and extend approximately from 4 to 16 kilometers, or more than twice the depth of the surface trades. Still higher and above the counter-trades flows the so-called upper easterly current, extending up to a height of 20 kilometers, and above this again, a westerly flow in the same direction as the counter-trades, and approximately 5 kilometers in depth. Finally at a height of 30 kilometers there would seem to be another easterly current. Thus over the tropics we find at least five wind systems. As we move to higher latitudes, the winds, possibly

Anti-trades and counter-trades

Upper easterlies

under the influence of the deflective tendency due to the earth's rotation, change their direction through the south and become eventually west winds. These air streams are drier and heavier than the trades, and descend to the surface in latitude 30° from the thermal equator, as warm, dry southwest winds. In the United States, the antitrades are more commonly called the prevailing westerlies, winds which lack the steadiness of the trades, but which nevertheless are the controlling factors in determining the weather of the temperate zones.

Prevailing westerlies

In connection with the movement of the upper air it is of interest to note that the dust from the Krakatau eruption in 1883, a few degrees south of the equator, was carried from east to west around the world in about 15 days. The red sunsets[1] and sunrises due to the fine dust and vapor particles appeared progressively later from west to east and indicated an average movement of 113 kilometers per hour, 31 m/s.

28. Monsoons. The word *monsoon*[2] is said to be of Arabic origin, meaning "season," and rightly applies to the winds of the Indian Ocean, for the general character of the season and the crop yield are closely connected with the duration and intensity of these winds. During the summer months the southwest monsoon, force 3 to 5 (3 to 8 m. p. s.), dominates the ocean north of the equator. It overspreads the Arabian Sea early in June, and by the third week is in full force over the Bay of Bengal. Severe thunderstorms, thick, cloudy weather, and gales with occasional dangerous cyclones occur during the period immediately preceding the full force of the monsoon. In winter we have to deal with the northeast monsoon, force 3 to 4 (3 to 5 m. p. s.), which prevails over Indian waters and extends down the African coast to latitude 10° S. Northwesterly winds prevail in the Persian Gulf and the Gulf of Oman, and easterly winds in the Gulf of Aden. In the southern part of the Red Sea the winds are southeasterly, and in the northern part they are

The summer monsoons

The winter monsoons

[1] See end of section 44, "Why sunsets are red."

[2] Alexander the Great is said to have brought back to Greece, after his invasion of India, information concerning the monsoon.

northwesterly. On the Asiatic coast the winds east of Chosen (Korea) are northeasterly; west of it they are northwesterly. Along the China coast immediately north of Shanghai to the 5th parallel they are northeasterly and are known as the northeast (winter) monsoon. The monsoon is in full force during January, and blows with greatest strength and constancy between Macao and Chusan. It shows a marked tendency to follow the coast, and as it weakens at night and the wind becomes somewhat offshore, northbound sailing vessels may then make fair headway. The thick, rainy weather of the monsoon period renders navigation difficult off the coast of Taiwan (Formosa). A rising pressure foreruns an increase in the strength of the monsoon, and a falling pressure a decrease.

29. Local winds. In nearly every land there are local names for special winds, based as a rule upon the warm or cold, and wet or dry character of the wind. In mountainous countries, especially if the range is but a short distance from some large water surface, the air at times seems to rush through the valleys and cañons. This can nearly always be traced to the passage of some general disturbance. There is another class of day-and-night winds which are due primarily to differences of temperature in the valley and at the level of the mountain tops, also sometimes to differences in the heating of the east and the west sides of the range. These are the well-known mountain and valley winds, reversing their direction with the change from night to day. In all these wind systems the contour of the land plays an important part. Study of the topography shows that the drafts are localized and intensified by the lay of the land. Most of these winds are in the nature of forced drafts, in the sense that air masses, generally with moderate momentum, forced through restricted channels, such as mountain passes and valleys, are drafts. These are chiefly horizontal currents, while the regular mountain and valley winds are more often due to vertical currents.

Winds influenced by topography

It is not surprising, therefore, to find that the direction of the flow may be determined to some degree by the trend of the narrow air passage or valley. Thus, although the foehn

wind in the Alps is primarily a south or southwest wind, it may appear in certain districts as a southeast wind, having had its direction of flow deflected by the trend of the valley. Furthermore, displacement at one place means motion at some other point of the circuit, and we may have an endless chain, as it were, in which the natural flow is masked. And just as in the case of the flow of water in rivers there may be established return currents at the sides of the main stream, or eddy currents at points where obstruction to the general flow is met, in the central part of a valley the flow of air or wind may be in one direction, while on the sides the flow may be in an opposite direction. An excellent way of studying the flow of air in mountainous countries is from a station on the summit. Close observation of the clouds above and the fogs below, as they form and dissipate, will show the existence of many unsuspected air streams.

Forced drafts of mountain passes

Of all the special winds, the foehn is perhaps the best known as it has been most studied and written about. The word is of German origin and possibly is connected with the Latin *favonius*, a west wind of the spring; but if such were the original meaning it is not in accord with the conditions now existing, for the foehn is essentially a dry south wind. It blows on the northern slopes of the Alps and is most noticeable in those valleys which have a north-and-south trend; indeed, it is hardly noticeable in some valleys which extend east and west. For many years it was explained as originating in the deserts of Africa; but it is now known to be the southerly component of an indraft due to the passage of a cyclonic area over Western Europe. The word *foehn* is also used in a broader sense to designate any wind system where the air, moving into a cyclone, is forced over some range and thus cooled and dried; and then, descending on the farther slopes, is dynamically heated. Under such conditions evaporation is rapid and snow on the ground disappears quickly. In the Northern Hemisphere such winds in temperate latitudes are generally from the south, and in the Southern Hemisphere

The foehn wind

Place of origin

The foehn type of wind

from the north. Thus we find foehn winds in Greenland, Iceland, Eastern Europe as in Hungary (*rotenturmwind*), South America, Japan, Peru, and in fact in all parts of the world where mountains act as partial barriers to the flow of air and there is compression after expansion. Such a wind is the

The chinook

chinook (name of an Indian tribe dwelling on Puget Sound), a dry and relatively warm wind of Wyoming, Montana, Idaho, eastern Oregon, and parts of Colorado. A good illustration of the pressure distribution and resulting wind direction and temperature can be found on the weather map for January 23, 1907. The temperature rose quickly from about 255°A. to 275°A. and, as previously stated, the snow evaporated rapidly. The duration of the wind depends upon the movement of the low-pressure area to the north. Sometimes the high temperature will last twenty-four hours. Other warm, dry winds are

"Hot winds," "northers," "Santa Anas"

the so-called "hot winds" of the Plains states, the summer winds of Texas, the "northers" of the Sacramento and San Joaquin valleys, and the "Santa Anas" of southern California. Some of these winds in the summer months pass over heated areas and are warmed to some degree by radiation from the earth. The

Sirocco

sirocco of southern Italy and Greece is a warm south wind, generally dust-laden and therefore trying to man and beast. The *leveche* of Spain and the *leste* of the Madeira Islands are sirocco winds; the *solano* is an east wind on the east coast of Spain; the *harmattan* is a hot, dusty east wind of the winter months in the Gulf of Guinea; the

Harmattan and simoon

simoon (from the Arabic word for poison, although there are no poisonous gases associated with it) is a hot, sand-laden wind felt in Palestine, Syria, and Arabia; the *khamsin* of Egypt is a hot southeast wind which blows for about fifty days after the middle of March; the *brickfielders* are hot north winds of Southern Australia. There are many others having local names.

30. Cold waves and boreal winds. The word *boreal*, from *Boreas*, the north wind of the Greeks, is used to designate a class of cold winds generally of cyclonic origin. There would seem to be some connection between the intensity of

the depression and the temperature of the northwest quadrant. Being preceded by comparatively warm southerly winds, the contrast is marked and all the more noticeable. The air is not necessarily brought from high levels, and the compression is not sufficiently great to warm the air enough to affect materially its initial low temperature. The so-called cold waves of the United States are essentially boreal winds. Such, too, are the *buran* or *purga* of Russia, the *pamperos* of Argentina, the southerly *burster* of New Zealand, the *bora* of the Adriatic, and the *mistral* of the valley of the Rhone. The word "mistral" is derived from the Latin *magister*, and the wind is therefore appropriately described as a master wind, or the wind which dominates. It has been known for some time that the winds of the Antarctic region were of higher velocity and lower temperature than elsewhere, and the records of the Australian Antarctic Expedition of 1911–1914 confirm this. Thus gusts of wind having a velocity of nearly 90 meters per second (200 miles per hour) were recorded on the Robinson anemometers used. The record is subject to correction, and these figures may be reduced 20 per cent. Velocities of 80 meters per second and even higher were not infrequent, nor were winds of 45 meters per second (100 miles per hour) with a temperature of 240°A. (−28°F.) rare.

31. Charts for aëronauts and aviators. The term "aëronaut" is used to designate the pilot of a balloon, while "aviator" is restricted to the pilot of a heavier-than-air flying machine. One of the first attempts to bring the results of the exploration of the air by kites, balloons, and other means into convenient form for the use of aviators and aëronauts is the volume by Rotch and Palmer,[1] issued in 1911. Charts of the relative heights, corresponding densities, and temperatures are given. The frequency of winds from various directions and their respective velocities at Blue Hill are shown. Thus the shallowness of easterly winds is made evident by comparisons at different levels. The summer sea breeze has a depth of about 1,000 meters, while the easterly winds of cyclonic origin may have a depth of 2,000 meters. The winds of winter

Shallowness and depth of winds

[1] *Charts of the Atmosphere for Aëronauts and Aviators.*

are of greater velocity than the winds of summer. In brief west winds are most frequent. Near the ground they blow about 25 per cent of the time from south-southwest to west-southwest in summer, and about 33 per cent of the time from west to northwest in winter, with a velocity varying from 8 to 11 meters a second. At a height of 3,000 meters the frequency of the westerly winds increases to nearly twice that at the lower level, and there is a corresponding increase in velocity. Particular attention is paid to the problem of trans-Atlantic flight and the possibility of utilizing the northeast trade for the western voyage is considered. The height at which the southwest counter trade may be expected is uncertain. It has sometimes been found below 1,500 meters, and again has been absent at 10,000 meters.

Summer and winter variations

The trades and trans-Atlantic flight

Another excellent book is that of C. J. P. Cave, entitled *The Structure of the Atmosphere in Clear Weather*, which gives the result of two hundred observations of pilot balloons and *ballons-sondes*. The direction and velocity of the wind at different levels are charted. Cardboard models show the distribution of wind with each kilometer of height. The arrowhead flies with the wind. The gradient wind is computed from the distribution of pressure. The velocity is calculated from the measured distance of two isobars between which the station lies by means of the formula

Cave's models

$$G=2\omega\rho V \sin \phi,$$

where G is the gradient, ω the angular velocity of the earth, ϕ the latitude, V the velocity, and ρ the density of the air.

Five types of atmospheric structure are described: (1) where the wind in the upper air is steady and there is no increase of velocity with height; (2) where the wind is steady but increases in velocity much above the gradient value; (3) where the upper wind decreases in velocity; (4) where changes of direction or reversals occur; and (5) where the upper air blows away from centers of low pressure.

Five types of atmospheric structure

The strongest current is, as a rule, just below the stratosphere; and in view of the work of Dines and the suggestions of Shaw, the question is raised whether it would not be advantageous to refer variations in the different levels to the conditions in the 9-kilometer level instead of to the surface. Starting with a strong westerly wind under the stratosphere, Cave would then work downward, for almost without exception the west wind decreases in the lower levels and the falling off may proceed continuously to such an extent that the direction of motion is reversed at some point in the intermediate layers, so that near the surface an easterly wind is shown instead of the westerly one of the upper levels.

Strength of air current varies with height

The term "holes in the air" has been used by aviators to describe certain unstable conditions experienced when flying and which in their opinion are caused either by "holes" in the air, or by partial vacuums or "pockets." Such conditions are found on summer mornings and afternoons and near cumulus clouds, and are generally simply ascending currents of some momentum through which the flyer passes. Sometimes the aviator may skirt such a column of uprising air (and there are also descending currents) and part of the plane be within the current while the rest may be without. In such cases there will be sudden tilting or inequality of pressure, and the aviator should be careful not to attempt to meet the changed conditions too quickly, for they are but temporary and instability will result when the machine is again free. Not only near cumuli but also near cross currents, — that is, where one air stream is flowing in a different direction from an adjacent stream,— there will be minor vortices and more or less of an air surge; and such a condition will cause instability of the aëroplane. As has been explained in preceding paragraphs, there is sometimes a marked stratification of the lower air, and under certain conditions marked turbulence. When such conditions are suspected it might be advisable to resort to preliminary tests by freeing pilot balloons or pilot planes.

So-called "holes in the air"

Ascending and descending currents

CHAPTER XI

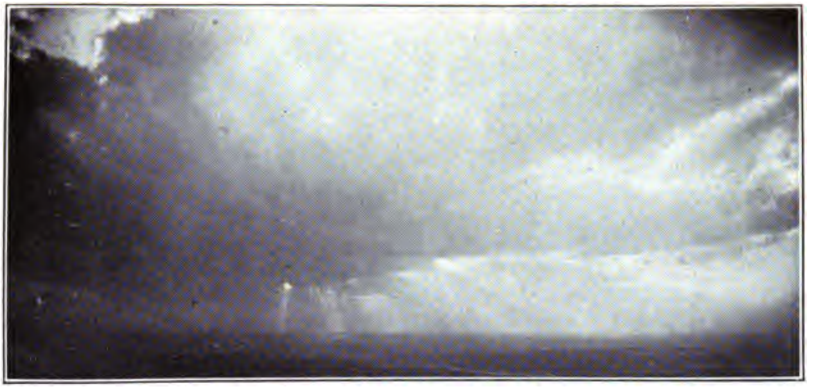
THE WATER VAPOR OF THE ATMOSPHERE

32. Earliest knowledge of cloud formations. We may well wonder what primitive man must have thought of the clouds. Very likely in the beginning he tried to escape from them, thinking they were in pursuit of him. Both the cloud and the cloud shadow must have seemed to him animate bodies; and so he must have watched them with awe, knowing that they came noiselessly and could outrun him. Some of them, the dark storm clouds in particular, must have terrified him beyond measure, and the hail, the lightning, the thunder, and the rush of the wind must have seemed to him the movements of an enemy—an angry cloud spirit—bent on doing harm. As time passed familiarity with nature lessened this fear of man for the clouds, and after a time he paid little heed to them, except perhaps when at sunrise and at sunset the rare forms and exquisite coloring appealed to some religious sense and called forth his admiration, and even a feeling of worship.

Effect of clouds on primitive man

Although many years have passed since men first marveled at the clouds, we still know little about them. True, we have lost our fear; and most persons understand that these "nurslings of the sky" are but forms and shapes of condensed water vapor; yet how little do we know of the whole story of the birth, life, and death of these ever-changing forms. In every wandering cloud there is a wonder tale of forces operating at different levels. We know much about the expansive power of water vapor when harnessed in the steam engine, and all that this force means for the welfare of men; yet a similar agency, the expansive power of water vapor operating in the free air, has received comparatively little attention.

From the ancients we get nothing of real value in connection with a knowledge of clouds. Nor do we get any important knowledge from the philosophers of the sixteenth



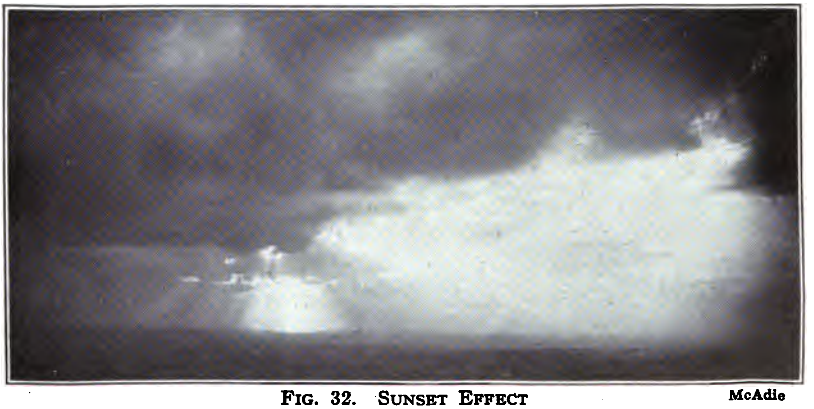


FIG. 32. SUNSET EFFECT McAdie

FIG. 33. SUNSET EFFECT, THIRTY SECONDS LATER THAN FIG. 32

FIG. 34. SUNSET EFFECT, THIRTY SECONDS LATER THAN FIG. 33

and seventeenth centuries. This is rather surprising, considering the general interest then shown in the physical sciences.

33. Measurements of cloud altitudes. We must note, however, that the height of certain kinds of clouds was measured trigonometrically from two stations as early as 1644 by Riccioli and Grimaldi, Jesuit priests of Bologna. Notwithstanding this, and in spite of the general awakening of interest in air exploration, there were no definite measurements of cloud altitudes until 1884, when Eckholm and Hagström systematically measured cloud heights at Upsala. About the same time attempts were made to employ photography in cloud measurements at the Kew Observatory. Probably the first trigonometrical cloud measurements made in America (rendered possible by the use of a telephone between the two observing stations) were those of W. M. Davis and A. McAdie, at Harvard University in 1885. In 1890–1891 the methods used by the Swedish investigators were employed by A. L. Rotch at Blue Hill Observatory, and for several years elaborate measurements were made by Clayton, Fergusson, Sweetland, and others.

Riccioli, Grimaldi

Photography employed in cloud measurement

The *Discussion of the Cloud Measurements* by Clayton is one of the best contributions to our knowledge of cloud formation and motion yet published. Other elaborate discussions are those of Sprung and Süring, which explain in detail the cloud observations made at Potsdam for the year 1896–1897; and those contained in the report of Professor Bigelow on the international cloud observations, published by the United States Weather Bureau in 1900. Professor Bigelow's report includes, besides observations on cloud heights, discussion of typical storm circulations and the relation of cloud motions to cyclones and anticyclones.

Clayton

34. Classification of clouds. An excellent "popular" book on cloud studies is that by A. W. Clayden, published in 1905. Reference will be made later to the work of Hildebrandsson, Ley, and others who have made special investigations of cloud types. Lamarck, the great naturalist, published in 1801, in the *Annuaire de la Société Météorologique* of Paris, a

classification of the clouds based upon their appearance. This, of course, is not the correct basis for classification, but not until recent years has there been any proposal to classify clouds on other and more scientific principles. Lamarck divided the clouds into small groups: those resembling somewhat flocks of sheep; large cloud masses; extended thin layers, or sheets; and piled-up masses, or large individual clouds. In all he described about twenty different and easily recognizable cloud forms. The first attempt at cloud classification in English was made by Luke Howard, a young chemist of Tottenham, who read before the Askesian Society, session of 1802–1803, an essay in which he proposed certain names. The essay was published in the *Philosophical Magazine* for 1803. Beginning with the lowest clouds, Howard proposed the term *nimbus* for rain clouds; *stratus*, or "layer," for the widespread, flat formations; *cumulus* for the rounded, piled-up forms; and *cirrus* for the high, feathery, or wisp-like types. By combining these, all ordinary forms could be accounted for. Howard's system was so flexible and so easy of comprehension that it met with favor and general acceptance. The essay was reprinted in 1832, translated into various languages, and adopted almost without change by the various official meteorological bodies throughout the world. Howard was much honored for this and other work. But the Howardian classification is not a scientific one, since it is based upon appearances; and we now know how misleading appearances may be. There is need of a classification based upon cloud origin. In aërography the cloud is significant, as it tells of the physical processes operative in the free air. It should tell of the motion of the air, the nucleation, and the thermal energy involved. The cloud is an exponent of the rate of condensation. It represents a definite problem in the absorption and radiation of energy, and also is an agency for the transportation of energy. Howard's system is not sufficient to meet the needs of our times, and ultimately there must come a classification suitable for the needs of modern science. There have been many investigators since Howard, representing many nationalities,

Lamarck's classification of clouds

Howard's classification

Why clouds are significant

but few of them have proposed entirely new systems, and nearly all have been content to modify slightly the types proposed by Howard. The list includes Poëy, Forster, Clos, Kaemtz, Fritsche, Jevons, Clouston, Mühry, Ley, Weilbach, Vettin, Klein, Köppen, Tissandier, Barker, Möller, Toynbee, Jesse, Abercromby, Hildebrandsson, Maze, Singer, Neumayer, Kassner, Riggenbach, Akerblom, de Bort, Fitzroy, Clayden, Clayton, Gaster, Vincent, and many others.[1]

A remarkable cloud record is that of S. C. Russell, extending over eight years. He accumulated in the course of this period nearly a hundred thousand observations.[2]

Ley's classification

In 1894 Clement Ley, in his book entitled *Cloudland*, proposed a new classification consisting of four main divisions: (1) radiation clouds, (2) interfret clouds, (3) inversion clouds, (4) inclination clouds. Under the first division are all the fog types; under the second, clouds caused by the interaction of horizontal currents; under the third, the cumulus types, or clouds caused by condensation due to convectional currents; and under the fourth, the cirrus types.

Clayton's classification

Clayton, in 1889, suggested the following classification[3] based upon the *origin* of the cloud:

1. Clouds due to local, nearly vertical, ascending currents. In this group belong the forms known as cumuli.
2. Clouds due to slow and oblique ascending currents. In this group belong the sheet clouds of stratification.
3. Clouds due to the chilling of the lower air by radiation from the earth. In this class belong the fogs.
4. Clouds due to the evaporation of the thinner parts of clouds already formed, probably caused by descent. In this group belong many of the clouds which appear in "flocks" of balls or rolls; also certain forms of cirrus.
5. Clouds due to differences in the direction and velocity of air currents at different levels. In this class belong the cirrus.

[1] Goethe was much interested in meteorological phenomena, especially in cloud forms. He carried on a correspondence with Howard, which extended over many years. An interesting statement of Goethe's meteorological views, together with nine sketches of cloud types, is given in W. v. Wasielewski's *Goethes meteorologische Studien*, Leipzig, 1910.

[2] *Quart. Jour. of the Royal Met. Soc.*, Oct., 1913. The clouds are grouped in four main classes: (1) upper clouds; (2) intermediate clouds; (3) lower clouds, including fog; and (4) diurnal ascending current formations.

[3] *Annals Harvard Observatory*, Vol. XXX, Part IV, 1896.

About 1890, chiefly through the efforts of Hildebrandsson, an international cloud classification was agreed upon, and in

McAdie

FIG. 35. MORNING FOG RISING. SEA FOG AUGMENTED BY RADIATION FOG

1896 there appeared an international cloud atlas, by Hildebrandsson, Riggenbach, and de Bort.

35. The international system. During the summer of 1894 a committee, which had been appointed by the International Meteorological Congress at Munich, met at Upsala to prepare an atlas representing the cloud forms with the nomenclature proposed by Hildebrandsson and Abercromby, and recommended for general use by the congress. The definitions adopted by this committee are as follows:

1. *Cirrus* (Ci.). Isolated feathery clouds of fine fibrous texture, generally of a white color, frequently arranged in bands which spread like the meridians on a celestial globe over a part of the sky and converge in perspective toward one or two opposite points of the horizon. In the formation of such bands Ci.S. and Ci.Cu. often take part.
2. *Cirro-stratus* (Ci.S.). Fine whitish veil, sometimes quite diffuse, giving a whitish appearance to the sky, and called by many "cirrus haze," sometimes of more or less distinct structure, exhibiting tangled fibers. The veil often produces halos around the sun and moon.

FIG. 36. STRATO-CUMULUS. LOW FOG McAdie

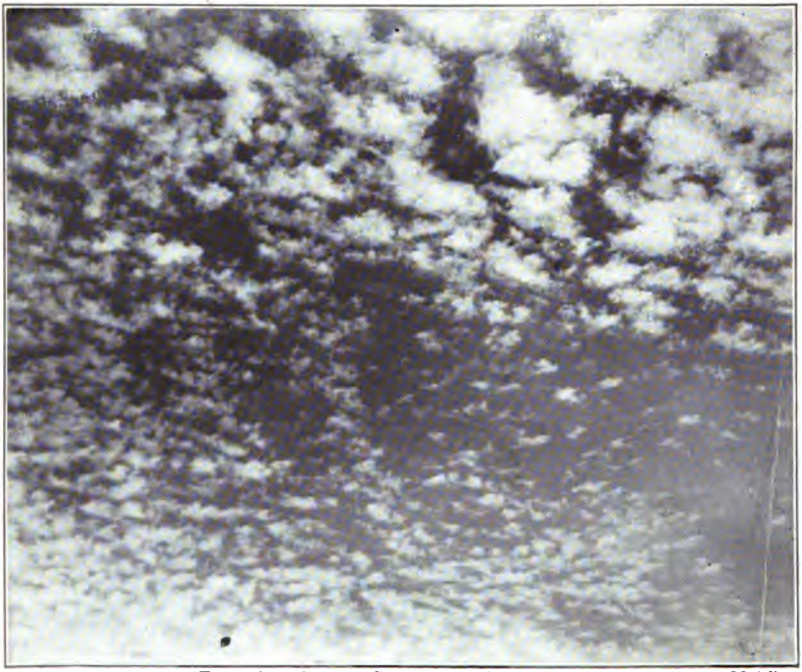


FIG. 37. CLOUD CHANGES. ALTO-CUMULI McAdie

3. *Cirro-cumulus* (Ci.Cu.). Fleecy cloud. Small white balls and wisps without shadows, or with very faint shadows, which are arranged in groups and often in rows.
4. *Alto-cumulus* (A.Cu.). Dense fleecy cloud. Larger whitish or grayish balls with shaded portions grouped in flocks or rows, frequently so close together that their edges meet. The different balls are generally larger and more compact (passing into S.Cu.) toward the center of the group and more delicate and wispy (passing into Ci.Cu.) on its edges. They are very frequently arranged in lines in one or two directions.

   The term "cumulo-cirrus" is given up because it causes confusion.
5. *Alto-stratus* (A.S.). Thick veil of a gray or bluish color, exhibiting in the vicinity of the sun and moon a brighter portion, which, without causing halos, may produce coronas. This form shows gradual transitions to cirro-stratus, but according to the measurements made at Upsala it has only half the altitude.

   The term "stratus-cirrus" is abandoned because it gives rise to confusion.
6. *Strato-cumulus* (S.Cu.). Large balls or rolls of dark cloud which frequently cover the whole sky, especially in winter, and give it at times an undulated appearance. The stratum of strato-cumulus is usually not very thick, and blue sky often appears in the breaks through it. Between this form and the alto-cumulus all possible gradations are found. It is distinguished from nimbus by the ball-like or rolled form, and because it does not tend to bring rain.
7. *Nimbus* (N.). Rain clouds. Dense masses of dark, formless clouds with ragged edges, from which generally continuous rain or snow is falling. Through the breaks in these clouds is almost always seen a high sheet of cirro-stratus or alto-stratus. If the mass of nimbus is torn up into small patches, or if low fragments of cloud are floating much below a great nimbus, they may be called "fracto-nimbus," the "scud" of the sailors.
8. *Cumulus* (Cu.). Wool-pack clouds. Thick clouds whose summits are domes with protuberances, but whose bases are flat. These clouds appear to form in a diurnal ascensional movement, which is almost always apparent. When the cloud is opposite the sun, the surfaces which are usually seen by the observer are more brilliant than the edges of the protuberances. When the illumination comes from the side, this cloud shows a strong actual shadow; on the sunny side of the sky, however, it appears dark with bright edges. The true cumulus shows a sharp border above and below. It is often torn by strong winds, and the detached parts present continual changes ("fracto-cumulus").

FIG. 38. CLOUD CHANGES. ALTO-STRATUS McAdie

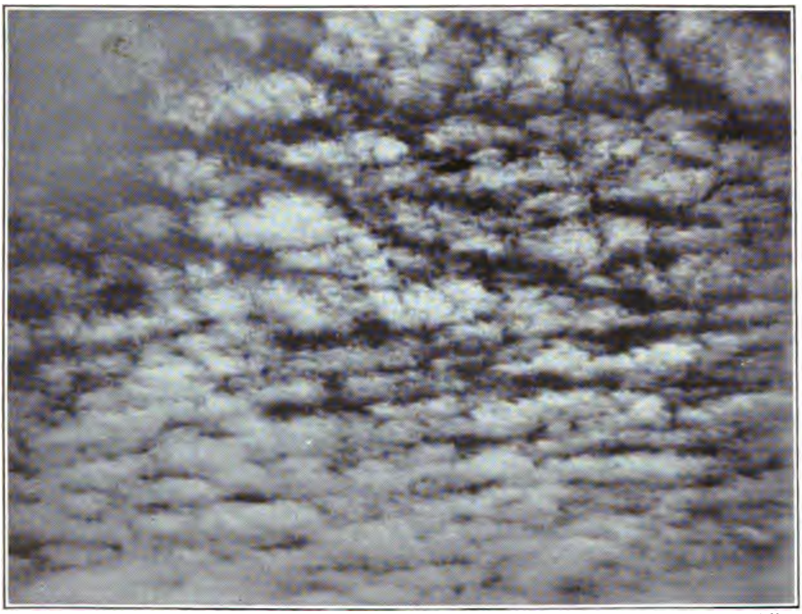


FIG. 39. CLOUD CHANGES. ALTO-STRATUS McAdie

Figs. 38 and 39 were taken at an interval of one minute.

9. *Cumulo-nimbus* (Cu.N.). Thunder cloud; shower cloud. Heavy masses of clouds, rising like mountains, towers, or anvils, generally surrounded at the top by a veil or screen of fibrous texture ("false cirrus") and below by nimbus-like masses of cloud. From their base generally fall local showers of rain or snow, and sometimes hail or sleet. The upper edges are either of compact cumulus-like outline, and form massive summits, surrounded by delicate false cirrus, or the edges themselves are drawn out into cirrus-like filaments. This last form is most common in "spring showers." The front of thunderstorm clouds of wide extent sometimes shows a great arch stretching across a portion of the sky, which is uniformly lighter in color.
10. *Stratus* (S.). Lifted fog in a horizontal stratum. When this stratum is torn by the wind or by mountain summits into irregular fragments, the clouds may be called "fracto-stratus."

The committee also adopted the following instructions for recording clouds:

"1. *The kind of cloud* designated by the international letters of the cloud name, which may be more exactly defined by giving the number of the picture in the atlas most nearly representing the observed form.

"2. *The direction from which the clouds come.* If the observer remains completely at rest during a few seconds, the motion of the clouds may easily be studied by noting their relative position to a steeple or other tall object, such as a mast, in an open space.

"If the motion of the cloud is very slow, for such an observation one's head must be supported. Clouds should be observed in this way only near the zenith; for if they are too far away from it, the perspective may cause errors. In this case, nephoscopes should be used, and the rules followed which apply to the particular instrument employed.

"3. *Radiant point of the upper clouds.* These clouds often appear in the form of fine parallel bands, which by an effect of perspective seem to come from one point of the horizon. The radiant point is that point where these bands, or their direction prolonged, meet the horizon. The position of this point on the horizon should be recorded in the same way as the wind direction, N., NNE., and so on.

"4. *Undulatory clouds.* It often happens that the clouds show regular parallel and equidistant striae, like the waves on the surface of water. This is the case for the greater part of the cirro-cumulus, strato-cumulus (roll-cumulus), and similar forms. It is important to note the direction of these striae. When there are apparently two distinct systems, as are to be seen in clouds separated into balls by streaks in two directions,

the directions of the two systems should be noted. As far as possible, observations should be made on streaks near the zenith to avoid effects of perspective.

"5. *Density and position of cirrus banks.* The upper clouds frequently take the form of a tangled web, or of a more or less dense veil, which, rising above the horizon, resembles a thin white or grayish bank. As this cloud form has an intimate relation to barometric depressions, it is important to note:

"(*a*) The density,—

0 meaning very thin and irregular.
1 meaning thin but regular.
2 meaning rather dense.
3 meaning dense.
4 meaning very dense and of dark color.

"(*b*) The direction in which the veil or bank appears densest.

"*Remarks.* All interesting details should be noted, for example:

"1. On summer days all low clouds generally assume particular forms more or less resembling cumulus. In this case there should be put under *Remarks*, 'Stratus or nimbus cumuliformis.'

"2. It sometimes happens that a cumulus has a mammillated lower surface. This appearance should be described by the name of 'mammato-cumulus.'

"3. It should always be noted whether the clouds appear stationary, or whether they have a very great velocity."

Changes in nomenclature

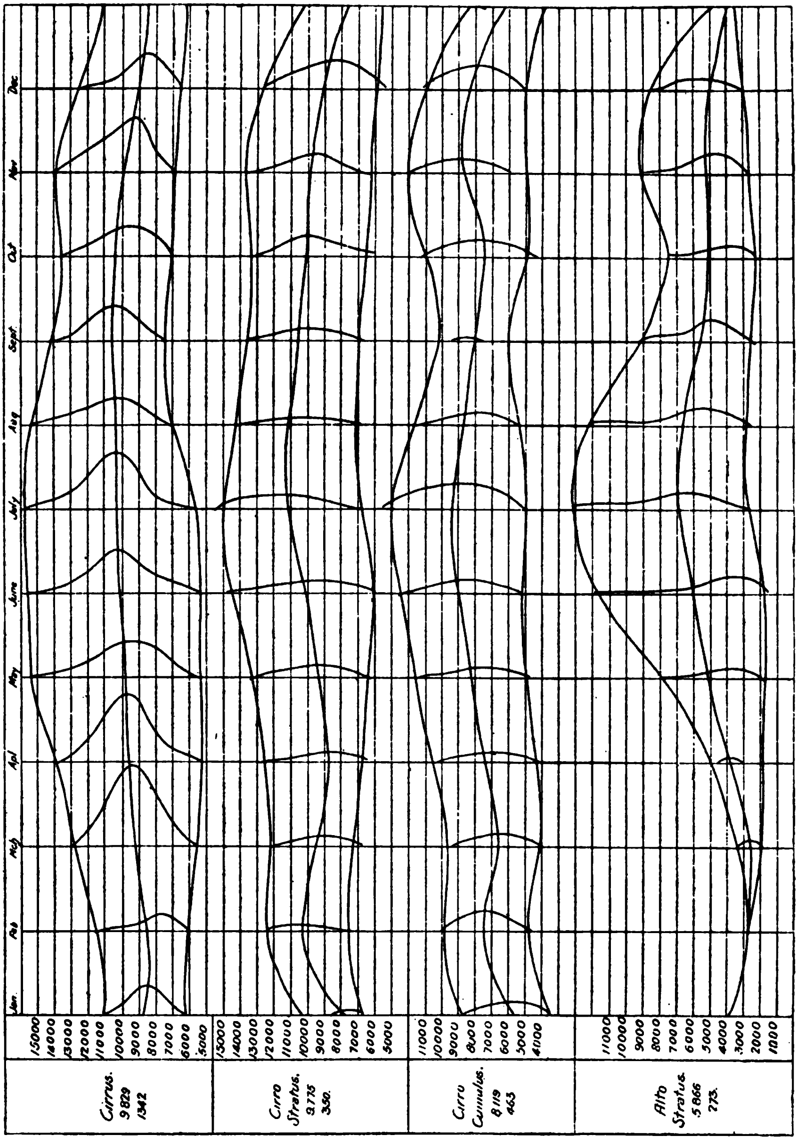
Clayton, in the *Discussion of the Cloud Observations*, says that "by following the changes in nomenclature since Howard, it seems clear that there has been a gradual evolution, during which differences and distinctions not recognized by Howard have been established, and errors due to perspective, as in the case of the cumulo-stratus, have been corrected. Thus distinctions between high and low cirro-stratus and between high and low cirro-cumulus have been established, and the lower forms called alto-stratus and alto-cumulus respectively. The stratus has been separated into fog and low sheet clouds, and two distinct forms of rain cloud are recognized. These distinctions have been a gradual growth, and Abercromby says: 'At Professor Hildebrandsson's suggestion we examined the nomenclature used by different offices, and arranged the names systematically; and we found that the differences did not seem irreconcilable. Eventually, we agreed that ten terms, all compounded of Howard's four fundamental types,— cirrus, stratus, cumulus, nimbus,— would fully meet the requirements

CLAYTON'S CLASSIFICATION OF CLOUDS ACCORDING TO ALTITUDE

Levels	Average altitude	Most frequent altitudes	Stratiforms	Cumuliforms	Flocciforms	Cirriforms
	meters	meters				
Stratus	500	600	Stratus (s) Nimbus (N) Fracto-stratus (fs) Fracto-nimbus (fN)	Cumulus *informis* (Ki)		
Cumulus	1600	1200 to 1700	Alto-nimbus (aN)	Cumulus (K) Cum-nim. (KN)	Nimbus *cumuliformis* (NK) Strato-cumulus (SK)	
Alto-cum......	3800	3000 4400	Alto-stratus *nimbiformis* (ASn) Alto-stratus (AS)		Alto-cum. (AK) Alto-cum. *tenuis* (AKT)	
Cirro-cum.....	6600	5800 7200	*Velo*-cirro-strat. (vcs) *Velo*-cirro-strat. (vcs)		Cirro-cumulus (CK) *Grano*-cirro-cum. (gCK)	
Cirrus	8900	8500 10000	Cirro-stratus (cs) *Lacto*-cirro-strat. (lcs)		*Flocci*-cirrus (flc)	Cirrus (c) Cirrus (c)

The letter "K" is used to indicate Cumulus.

The average altitudes given in the first column of figures for each level were determined by direct measurements. The *level* was recorded in each case from observation, and the altitudes were afterwards computed from angular measurements made at the same time with theodolites. This shows the possibility of distinguishing the five levels.



After Bigelow

FIG. 40. DISTRIBUTION OF CLOUD TYPES

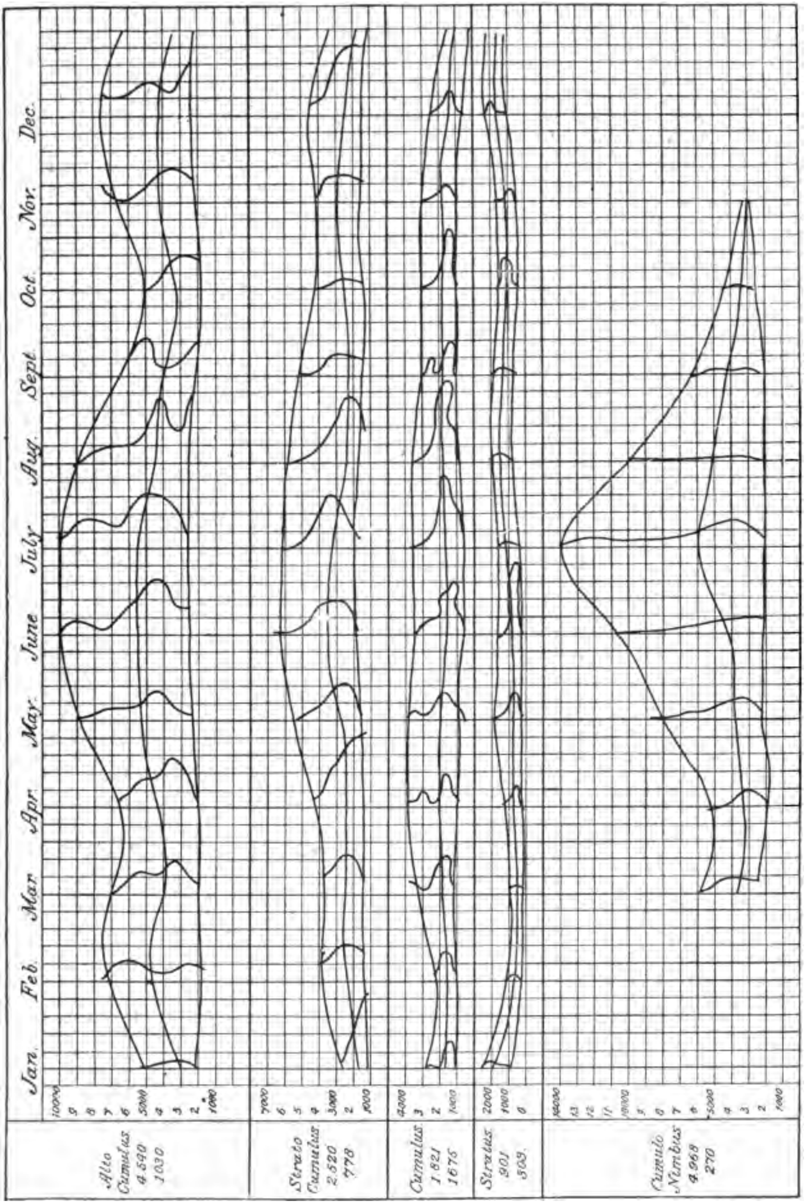
of practical meteorology, with the least disturbance of existing systems.'[1] Hildebrandsson further says that the ten cloud forms described were already recognized in the nomenclature used in Portugal. Hence the international cloud nomenclature adopted at Munich represents the greatest progress in cloud nomenclature which observers are yet ready to accept for general use; and no official bureau should hesitate to accept it for fear that the system is merely temporary and will soon be changed. Progressive development will undoubtedly continue, but changes of names in general use will, in all probability, be slow. A more detailed nomenclature is, however, needed for the use of specialists."

36. Distribution of the various types of clouds. Bigelow shows graphically (diagrams, Figs. 40 and 41) the distribution of the several types of clouds. Under the name of each type he gives the mean height in meters for the year and the number of observations. There is also plotted for the several types the curve of frequency, with heights as ordinates and the number of observations at the respective heights as abscissas. The curves follow the mean line of the plotted points very closely. Under the assumption that the observed frequency corresponds with the actual frequency of cirrus formation at the given height, a discussion of these curves would give a good explanation of the physical processes operative in cloud formation for the whole year.

In the diagrams the mean heights are shown, also the upper and lower limits. There is a wide range in the heights of certain clouds. The mean height of the low clouds is probably 2,000 meters. The three low cloud strata are shallow (not exceeding 3,000 meters in depth), except the cumulo-nimbus, or thunder head, which may develop a height of 13,000 meters. All the clouds except stratus and cumulo-nimbus show a tendency to three maxima of height and thickness, one in midsummer and the other two in February and November. The minima occur in March and September. A similar relation is found to exist in the isothermal limits, as pointed out in the chapter on the stratosphere.

Cirrus bands have been explained as due to differences in

[1] *Quart. Jour. of the Royal Met. Soc.*, April, 1887, p. 155.



After Bigelow

FIG. 41. DISTRIBUTION OF CLOUD TYPES

velocity or in direction of contiguous upper-air currents. These currents nearly always move from west to east, and the higher part of the current may move more rapidly than the lower. Thus the upper part of any cloud formation might move in advance of the base, causing a band or bar extending from west to east.

Cirrus bands

37. **Wave motions in the air shown by cloud undulations.** Cloud billows, or undulations, have a different origin from cirrus bands, though it is not always easy to distinguish between the two. According to Clayton, bands are usually isolated or widely separated and are of unequal length, while undulations are close, parallel rows or striations of nearly equal length. The undulations were comparatively little observed until Helmholtz called attention to them as illustrating wave motions in the air, of the same nature as ocean waves. These undulations are visible in clouds at all altitudes. They are illustrated in the strato-cumulus by long parallel rows, which are parallel in fact as well as in appearance, and lie in approximately the same direction in all parts of the sky. This can be seen by laying a ruler across the center of the mirror of the nephoscope parallel with the undulations. Abercromby had a different opinion, but he clearly made no critical observations in this way. The undulations are illustrated in the cumulus level by long parallel lines formed by individual cumuli like a file of soldiers, and the lines appear to converge toward the horizon as the effect of perspective. The undulations appear to be most frequent in the alto-cumulus level, and are easily distinguished by the parallel rolls in the alto-cumulus and by striations in the alto-stratus, like the furrows in plowed ground. In the cirrus level they are usually distinguished as short, parallel threads or as small bands forming one broad band at right angles to their length. Sometimes they seem like furrows in the cirro-stratus.

Cloud billows

The direction of length of the undulations is decidedly most frequent from north to south, which is at right angles to the most frequent direction of cirrus bands. It leads at once to the inference that cloud undulation is the phenomenon to which the popular name of "polar bands" was applied in Europe, but not to cirrus

"Polar bands"

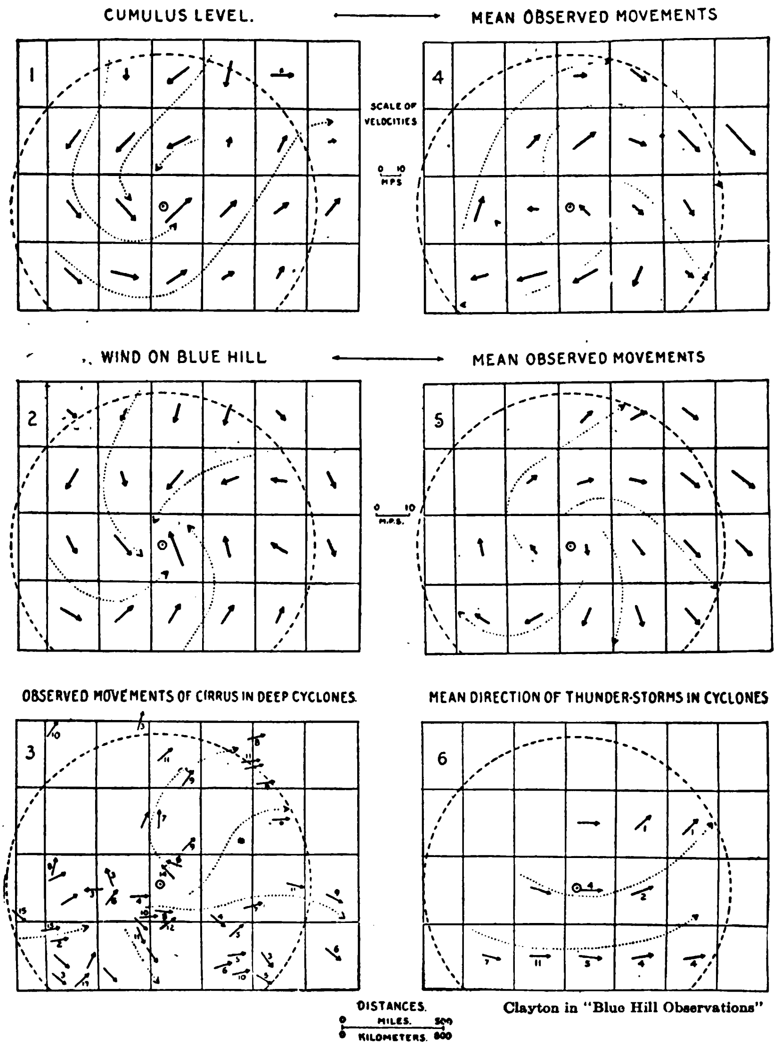


FIG. 42. DISTRIBUTION OF CLOUDS IN CYCLONES AND ANTICYCLONES

bands, as many meteorologists have supposed and have thus been led to introduce a wrong usage of the term. The tendency for the undulations to lie at right angles to their motion is very distinct, and is in contrast with the cirrus bands.

After Mohn

Fig. 43. Instrument for Measuring Cloud Heights

Since the crest of a wave usually lies at right angles to the wind which originates and drives it forward, it follows that the results of observation agree very well with Helmholtz's explanation of the cloud undulations; namely, that the clouds are the visible crests of real atmospheric waves formed between currents of air of a different density and having a different velocity or direction. The undulations would probably always lie at right angles to the upper current were it not that the lower current is also in motion, and that the observed cloud direction is a compound of the two.

Clouds as crests of atmospheric waves

38. The value of clouds in forecasting weather changes. The cloud is not, as might be expected at first thought, a good exponent of air motion; and as yet cloud maps have not been used advantageously by professional forecasters except in connection with storms of the West Indian hurricane type or the typhoon, or baguio, of the China Sea. Thus Father Viñes, at Havana, showed how certain types of cirrus accompanied or rather preceded storms of great violence but small diameter, while a different type was found to accompany storms of large

Clouds unreliable in forecasting

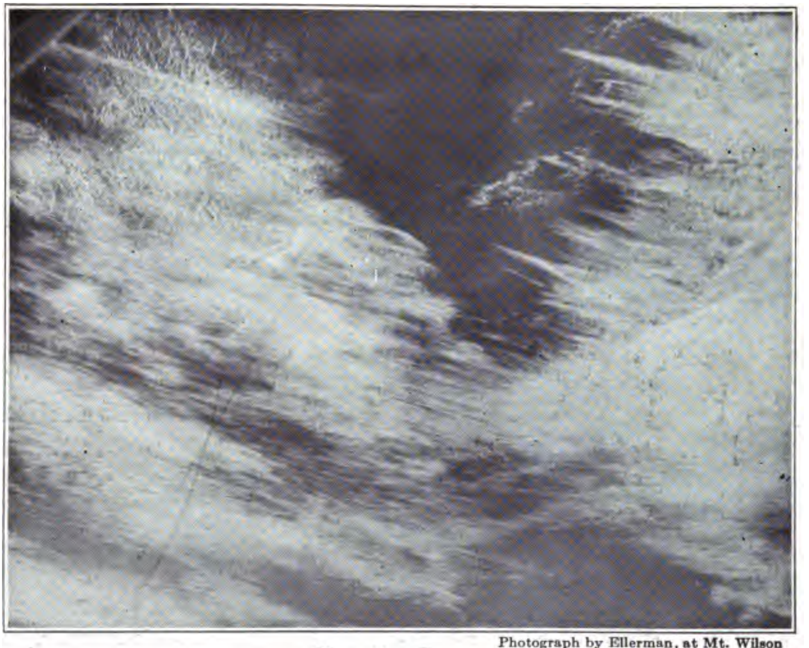
Clayton and Fergusson at Blue Hill

Fig. 44. Plotting Machine for Measuring Cloud Heights

Fig. 45. Edge of a Cumulus Cloud McAdie

diameter and moderate violence. Likewise at Manila, Zi-ka-wei, and other observatories on the Asiatic coast, the appearance and motion of the upper clouds have been carefully studied for forecasting purposes.

In general, cirrus clouds, except of a certain type, do not positively indicate coming rain, being, in fact, somewhat less frequently followed by rain than the average probability of rain. But they are closely controlled in their movements by

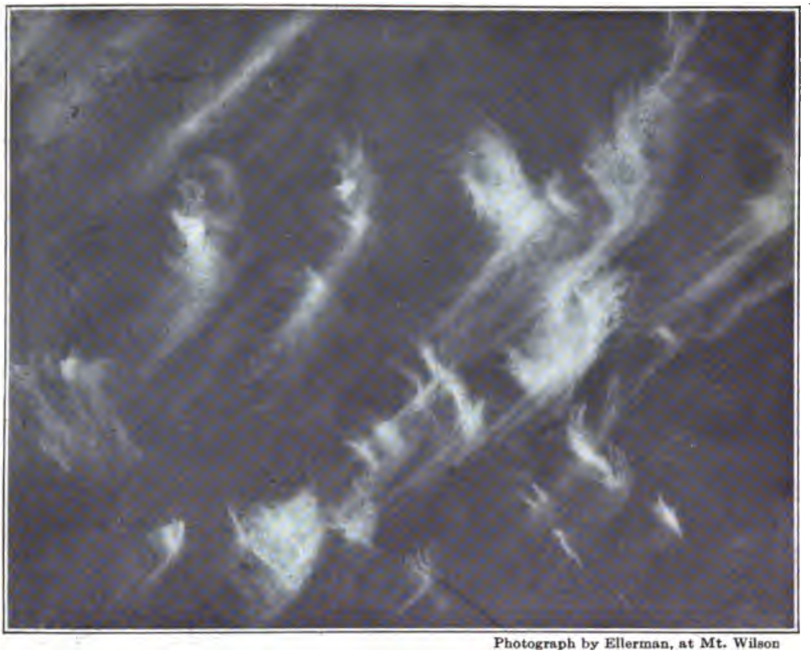


Photograph by Ellerman, at Mt. Wilson

FIG. 46. CIRRUS

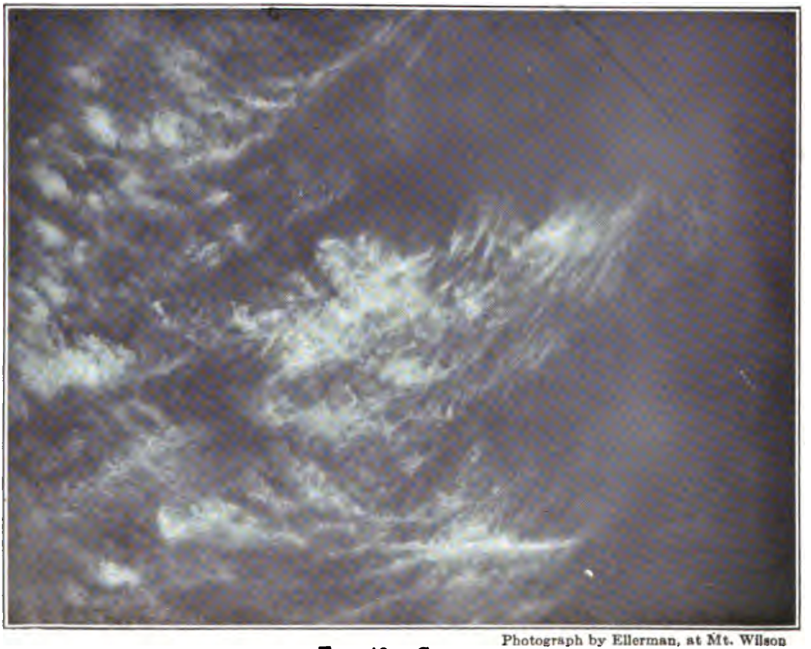
temperature gradients, and they may serve an isolated observer as a guide to coming changes of temperature. In general, slowly moving cirrus clouds indicate slight changes in temperature, and, except when moving from a direction between south and west, they indicate, as a rule, slowly rising temperature during the succeeding twelve and twenty-four hours. Rapidly moving cirrus indicate the probability of decided changes of temperature, and, from *any* direction, a

Cirrus clouds and forecasts of temperature



Photograph by Ellerman, at Mt. Wilson

FIG. 47. CIRRUS PLUMES



Photograph by Ellerman, at Mt. Wilson

FIG. 48. CIRRUS

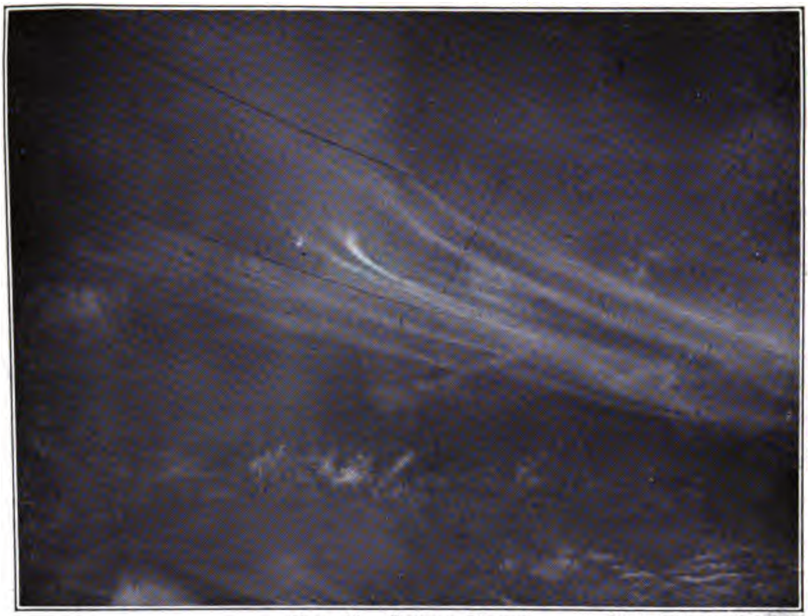


FIG. 49. CIRRUS BANDS Photograph by F. R. Ziel

probability of a fall of temperature by the end of twenty-four hours. The probability of a fall, however, and the amount of fall, are much greater and earlier when the cirrus are observed to be moving rapidly from a direction between south and west. When cirrus are observed to be moving from the southwest, there is a strong probability of a fall of temperature during the succeeding twenty-four hours. This probability rises to 83 per cent in winter and is over 70 per cent at all times of the year for cirrus moving rapidly from the southwest. When cirrus are observed to be moving from the northwest, the probability is that there will be a rise of temperature during the succeeding twelve hours. The probability is 64 per cent for winter and 76 per cent during the entire year for cirrus moving rapidly from the northwest.

Cirro-stratus and rain probability

With the appearance of cirro-stratus there is a probability of rain during the succeeding twenty-four hours of about 80 per cent. This probability increases to nearly 90 per cent with the appearance of alto-stratus, which is as great as can usually be derived from

FIG. 50. CLOUD FORMATIONS IN ADVANCE OF STORM McAdie

a knowledge of the conditions prevailing over the country as given on a weather map. Cirro-cumulus are most frequently followed by fair weather, while alto-cumulus indicate a probability of rain.

There are two directions in which observations of the direction and of the relative velocity of upper clouds might be of use. The rapid movement of cirrus from the west, or the southwest, along the northern boundary of the United States, will no doubt indicate the approach of a cold wave before its approach is indicated by the weather map, and will thus enable northwestern stations to be warned. The movement of cirrus from the south, observed at any of the Atlantic coast stations with a dense bank of clouds to the south of the observer, would strongly indicate a severe storm off the coast, and

When cirrus indicate coming cold

might enable the observer to determine the position of its center. This conclusion is derived from the individual observations at Blue Hill, and from the fact of circulation of air in cyclones with deep barometric depressions (Fig. 42). The prevalence of rapidly moving cirrus over a wide area indicates rapid storm movement, and rapid and marked changes of weather and temperature. Slowly moving cirrus indicate sluggish storm movements and slight changes of temperature, and are the usual accompaniment of droughts.

The direction of cirrus movement prevailing in advance of and around the storm center must, in the majority of cases, furnish a clew to the future movements of the storm, since it is found that the storm tends to move in the mean direction of the cirrus found for the storm as a whole.

39. **Recording sunshine.** For recording sunshine during the day hours various instruments are used. Some are

Photograph by Ellerman, at Mt. Wilson

FIG. 51. CUMULUS OVER MOUNTAINS

FIG. 52. CAMPBELL-STOKES SUNSHINE RECORDER

thermometric[1] and because of their responsiveness to heat, read too high if the temperature continues high after sunset, and too low if the temperature falls rapidly during the day; and some are photometric, based chiefly on the discoloration of sensitized paper. At Blue Hill there has been in constant use the well-known Campbell-Stokes sunshine recorder (Fig. 52). This record consists essentially of a glass sphere which focuses the sun's rays and burns a record upon a strip of prepared paper.

This instrument is also used to obtain a record of moonlight by placing a strip of photographic paper, not oversensitive, in the metallic frame, protecting it as much as possible from extraneous light and from the weather. Some very good records have been thus obtained, and it is quite easy to ascertain by these means the intensity of the illumination due to the moon. An interesting record (Fig. 53) is that of March 11–12, 1914, during a lunar eclipse. The moon entered the penumbra at 8:41 P.M. and entered the shadow at 9:42. The middle of the eclipse was at 11:13 P.M., the moon leaving the shadow at 12:44 A.M. and the penumbra at 1:45 A.M. The night was beautifully clear. On the record it will be seen that the light gradually became dim, disappearing about 11 o'clock and reappearing after midnight.

A record (Fig. 54) for the succeeding night shows that there was moonlight, with intervals of cloudiness, until 10:50 P.M., after which the moon was obscured. In the

[1] The Maring sunshine recorder used at Weather Bureau stations is essentially a differential air thermometer. Owing to greater heat absorption, the air in the black bulb expands more than the air in the bright bulb and the thread of mercury is forced upward, making an electrical circuit, when it meets the platinum tips. A circuit breaker in the clock of the quadruple register records minutes of sunshine. The instrument has a large twilight error, and must be frequently adjusted.

illustration the time scale has been enlarged to twice that of the other record.

Hours for recording state of the sky

The whole subject of cloudiness, or perhaps it would be better to say obscuration by clouds, is in a most unsatisfactory state. At most observatories it is considered sufficient to enter the state of the sky twice in twenty-four hours. In this country the hours are 8 A.M. and 8 P.M., but such a practice certainly does not afford a fair basis for estimating the relative cloudiness of the place. On the other hand, a systematic survey of the cloud forms requires endless time and patience.

One looks in vain through textbooks on meteorology, official circulars of instruction, and observers' handbooks for information regarding records of moonlight and night cloudiness. In fact, cloudiness during the dark hours, or practically half the time, is discreetly let alone by professional meteorologists. If recorded at all, the data are in abbreviated form, and are generally based upon hearsay, such as statements of night watchmen and others. Such data are, of course, of doubtful value. And yet many a case in criminal and civil courts requires evidence of a positive character regarding cloudiness at night, and particularly the illumination due to moonlight.

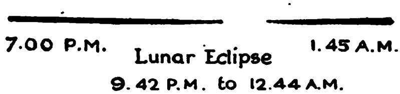


FIG. 53. MOONLIGHT RECORD DURING ECLIPSE

FIG. 54. MOONLIGHT RECORD WITH CLOUD INTERVALS

Night-cloudiness recorder

For getting a record of night cloudiness, there has been used at Blue Hill Observatory for many years a polestar recorder. This is a photographic record of a star trail, in this case the polestar, devised in 1885 by Professor E. C. Pickering. In 1904 the instrument was modified by Fergusson, and a record for two weeks may be obtained on one film; the instrument is practically automatic. The cost is not large, and there would seem to be no good reason why the instrument should not be generally used. If the night is clear, in the northern portion of the sky, the trail made by the star is continuous. For

example, in Fig. 55 the record for the night shows that the sky was entirely clear and not only the polestar but other star trails are continuous. Complete cloudiness would be indicated by absence of trail. In latitude 45° the instrument gives a fair record of night cloudiness for the entire sky.

While it is important, for many reasons, that cloudiness should be reported with some detail, still the records should be as compact as possible. A good illustration of a handy form of record is shown in Fig. 56. This was made for a lighting company. A similar record for the succeeding month showed an entirely different distribution of cloudiness; in fact, the month was so cloudy that the demand for light was double that of the preceding month. At Blue Hill the cloudiness for each hour throughout the year is charted on a sheet of millimeter paper 24 centimeters wide and 73 centimeters long. The record of cloudiness at Blue Hill is probably more complete than at any other point in the United States.

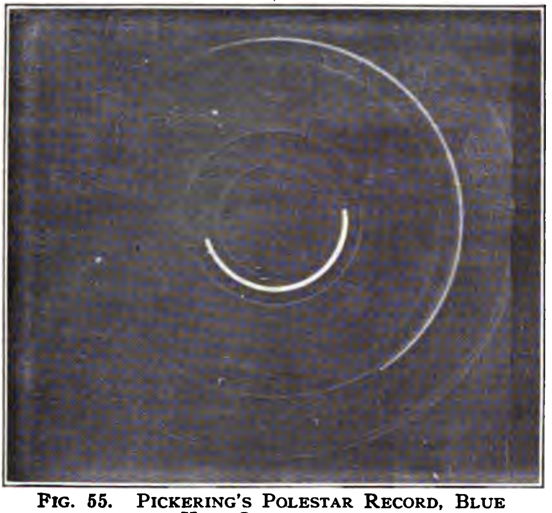


FIG. 55. PICKERING'S POLESTAR RECORD, BLUE HILL OBSERVATORY

An unusually interesting cloud record is that made at Epsom, Surrey, by S. C. Russell, who maintained an hourly record for eight years. During this period he accumulated nearly a hundred thousand individual records. He has published the results of monthly and hourly cloud-form frequencies in the *Quarterly Journal of the Royal Meteorological Society*, October, 1913. He groups the clouds into four main classes: (1) upper clouds, including all those of the cirrus type; (2) intermediate clouds, including most of the cumulus

and stratus formations; (3) lower clouds, including fog; and (4) diurnal ascending current formations. Curves showing hourly and monthly frequencies are given at some length. Nor is the matter of cloudless periods omitted, as is so often the case in cloud discussions. One remarkable cloudless period occurred in 1909, when from 6 P.M., April 4, until 5 P.M., April 11, a period of 167 hours, no clouds were visible. The duration of the condition of cloudlessness varies greatly with locality. On the Atlantic seaboard periods exceeding three days are rare; but in the far western part of the United States, especially in California and Arizona, periods

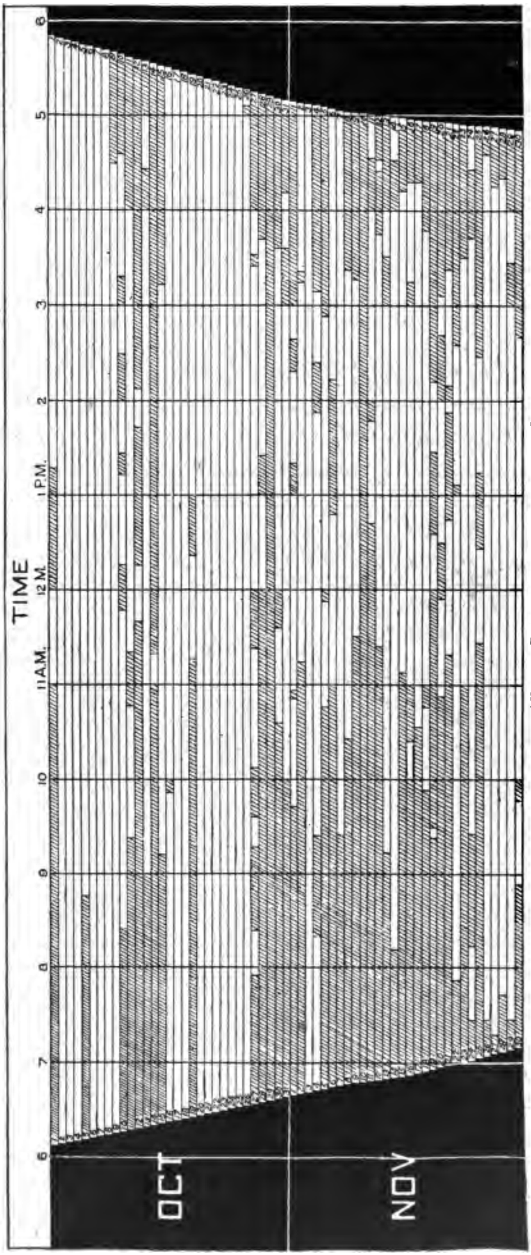


FIG. 56. RECORD OF DAY CLOUDINESS FOR TWO MONTHS

of a month or longer without cloudiness are not infrequent. Langley in his experiments at Mount Whitney speaks of the weeks which passed without a cloud in the sky; and it is common experience in the high Sierra during July, August, and September to find the sky entirely cloudless, week after week. On the other hand, in certain seasons, in these localities, thunderstorms may be frequent and afternoon cloudiness marked. The author once spent a week on the summit of Mount Whitney at the end of August, and more than half the time it was cloudy.

Long cloudless periods in the Sierra

Records of cloudiness are of especial interest to astronomers in connection with total eclipses of the sun. The whole purpose of an expedition to observe an eclipse of this nature may be defeated by a moment's cloudiness. When the path of totality is such that a choice of stations can be made, the astronomer gives preference to the locality having a record of minimum cloudiness.

CHAPTER XII

Condensation

40. The formation of clouds and the condensation of aqueous vapor. We have seen that raising a cubic meter of unsaturated air a little over 100 meters causes a fall in temperature of one degree. This has been called the "adiabatic rate," as the assumption is made that no heat is added or lost through other agencies. We have also seen that to raise the temperature of a cubic meter of *dry* air one degree requires 307 calories.

Adiabatic rate of cooling

When a mixture of air and vapor is lifted there will be expansion and cooling and, because of the cooling, a tendency toward condensation of the vapor. Under natural conditions heat may be added or lost through convection, radiation, and conduction. Unlike a solid, a gas when lifted or moved from a place of high pressure to a place where the pressure is less or the temperature higher, expands; work is done and heat expended—not in the lifting of the air or gas, but in overcoming both internal and external pressures. Work is also done in expansion when the mass of air is moved horizontally. In a perfectly homogeneous atmosphere, pressure and temperature would both decrease with height and offset each other in changing the density of the air. We should then have established a condition of unstable equilibrium. The air, if given a displacement upward, would continue to move upward, remaining always lighter than the adjacent air, the adiabatic cooling not being enough to lower the temperature to that of the high strata.

Expansion and cooling

Indifferent or adiabatic equilibrium occurs when the cooling brings the temperature to that of the new level. In this case no force will arise facilitating or opposing the displacement. But this condition very seldom occurs in nature, chiefly because of the effect of water vapor and the temperature change due to

When adiabatic equilibrium occurs

condensation or evaporation. As soon as condensation begins, the heat of condensation will partly offset the adiabatic cooling, and the adiabatic gradient will have such a value as that given in the following table from Bjerknes, which is a modification of the one given by Hann.

Adiabatic Fall of Temperature per 100 Dynamic Meters for Saturated Air

Pressure in kbs.	Temperature degrees absolute									
	263°	268°	273°	273°	278°	283°	288°	293°	298°	303°
300	0.52	0.46	0.40	0.42	.....	.....	.....	.....	.....	.....
400	.58	.52	.45	.47	0.42	.....	.....	.....	.....	.....
500	.63	.57	.49	.51	.46	0.41	0.37	.....	.....	.....
600	.67	.60	.54	.56	.50	.45	.40	.....	.....	.....
700	.70	.64	.57	.59	.54	.48	.42	0.39	.....	.....
800	.72	.66	.59	.61	.56	.50	.45	.41	0.38	....
900	.75	.69	.62	.64	.59	.53	.48	.44	.40	0.37
1000	.77	.70	.64	.66	.61	.55	.50	.45	.41	.38

While the adiabatic gradient for dry air is constant, that for saturated air varies with pressure and temperature, decreasing

Variation of adiabatic gradient

with pressure fall and increasing with temperature fall. As Bjerknes points out, the decreases upward both of pressure and of temperature counteract each other in their effect on the fall of temperature, making its variation with height gradual. But the gradient will increase upward, to the limit 1.0048, which would be reached when all moisture had fallen out. To illustrate this increasing fall of temperature, the values corresponding to the case of a mass of air with initial temperature 288°A. near sea level and moved upward, are shown by the underlined values in the table. It will also be noted in the table that two values are given for the freezing point as the transition is made from the solid to the liquid state. Practically, the value 0.5° for every 100 dynamic meters (101 meters) is used by aërologists as the average value of the temperature gradient for saturated air in the lower strata of the

Stable equilibrium

atmosphere. If temperature gradients are less than the adiabatic, we have a state of stable equilibrium. Under such conditions, if a mass of air be carried

up a certain distance, the adiabatic cooling will bring it to a lower temperature than that of the surrounding masses and it will fall again on account of its greater density. If there should be no gradient, we should have the density the same throughout; and the temperature at the highest level would be the same as below. This would be known as an "isothermic atmosphere." Considering extreme cases of unstable and of stable equilibrium, the first condition would occur if the fall of temperature amounted to 3.4° for every 100 meters. Such gradients may exist above a heated surface and probably for a short time over a heated area before the formation of a tornado. Contrasted with this is the case of a negative gradient, which is well known as a local phenomenon, under the name "inversion." It is a condition of pronounced stability requiring some effective source of heat to overcome it. It is important in connection with frosts and will be referred to later.

Inversion

Bjkernes groups the various states in four classes:

1. The homogeneous atmosphere, where, with a pressure of 100 kilobars, the temperature would be 27°A. and the gradient 3.48° per 100 dynamic meters.
2. The dry atmosphere in adiabatic equilibrium, where, with a pressure of 100 kilobars, the temperature would be 141°A. and the gradient 1°.
3. The ordinary saturated atmosphere, where, with a pressure of 100 kilobars, the temperature would be 196°A. and the gradient 0.5°.
4. The isothermic atmosphere, where, with any pressure, the temperature would remain the same, the gradient being zero.

These may also be arranged according to dynamic heights, in which case we should have at an elevation of 1,000 dynamic meters (1,000 ordinary meters are 980 dynamic meters) for

1, a pressure of 872 kilobars and temperature 238°A.;
2, a pressure of 878 kilobars and temperature 263°A.;
3, a pressure of 879 kilobars and temperature 268°A.;
4, a pressure of 880 kilobars and temperature 273°A.;

and at an elevation of 5,000 dynamic meters, for

1, a pressure of 362 kilobars and temperature 99°A., density same as at sea level;
2, a pressure of 494 kilobars and temperature 223°A., density 60 per cent of that at sea level;

3, a pressure of 512 kilobars and temperature 248°A., density 56 per cent of that at sea level;
4, a pressure of 528 kilobars and temperature 273°A., density 50 per cent of that at sea level.

It is not easy to apply adiabatic gradients to the actual condition, because the atmosphere is constantly gaining or losing heat through horizontal convection. Bigelow has deduced[1] the necessary corrections and auxiliary charts whereby the heat divergence between the assumed and the actual can be quickly obtained for purposes of forecasting. He also uses the temperature gradient in connection with cumulus clouds. The adiabatic gradient .98° A. per 100 meters is not often found in the lower strata. The bases of cumuli usually form higher than the surface values would indicate, and thus subheating is the rule, although superheating may occur on warm days near the ground. Starting at the surface with a given dewpoint and following on a Hertz or other diagram the line of constant saturation weight (Fig. 57), Bigelow deduces certain values which, with selected gradients, will hold throughout the given height, say the height of a cumulonimbus or thundercloud. He thus employs the lofty cloud as a gauge for obtaining temperatures at high levels.

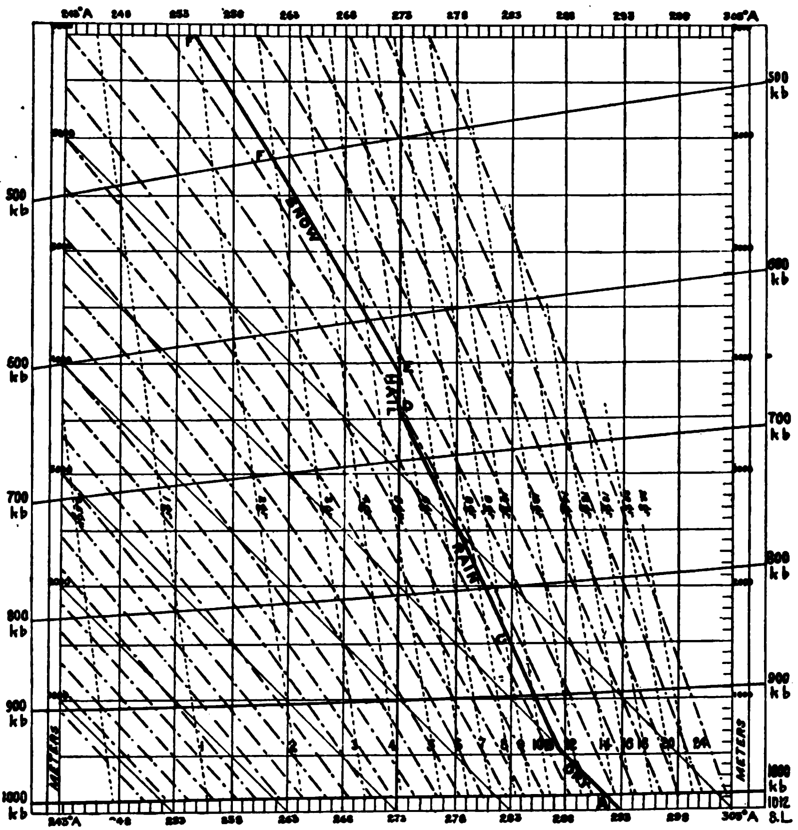
41. Conditions present in condensation. Condensation of the vapor into water decreases the rate of cooling with either elevation or lateral convection to lower pressure. If the condensation is carried farther and snow results, the gradient is still further lessened. In 1884 a brilliant young physicist, Hertz, expanded the small table for decrease of vapor with elevation, as given by Hann, and introduced a graphic method of following the changes in moist air. He discussed four stages: first, where the air is unsaturated and no liquid water is present; second, where the air is saturated and contains also additional fluid water; third, where, in addition to vapor and liquid, ice is present; and fourth, where there are only vapor and ice. The four stages are designated also as the dry, the rain, the hail, and the snow stage. It will frequently happen that the temperature down to which

Rate of cooling decreased by condensation

Four stages accompanying the changes in moist air

[1] *Report of the International Cloud Studies.* Also Bigelow *Atmospheric Circulation and Radiation.*

the first stage holds good will lie below the freezing point; in that case, one passes directly over to the fourth stage.



Modified from Neuhoff-Hertz

FIG. 57. ADIABATIC DIAGRAM

If we had to deal with only one mixture, whose composition was exactly known, and only one value of the ratio of the weight of the unsaturated aqueous vapor to the weight of the dry air, then we could represent pressure and temperature by coördinates in one plane and cover this with a system of curves connecting all those conditions which could adiabatically occur. But the aërographer must deal with mixtures of varying proportions. We can get along with one graphic table

if we confine ourselves to those cases in which the weight and pressure of the aqueous vapor are small in comparison with the air. One cannot expect too great accuracy. Then the same curve can be used for different mixtures; but the points at which the different stages pass into each other will be located differently; for this, special devices are needed. The solution of this problem was the origin of the Hertz diagram. Pressures are laid off as abscissas, and temperatures as ordinates. The diagram is so constructed that an equal increase of distance corresponds to an equal increase in the logarithm of the pressure and in the logarithm of the absolute pressure.

Methods of making adiabatic diagrams

In a paper presented in 1900, Neuhoff modified the Hertz diagram and gave both numerical and graphic methods for adiabatic changes of condition of moist air. A new diagram, showing the diminution of temperature with altitude, is so arranged that pressure is represented by cross lines with a slant and is read by scales on either side of the diagram. The stages are the same as those in the Hertz scheme. In the first the cooling proceeds without saturation, hence without precipitation; it is therefore called the *dry* stage. In the second, saturation occurs accompanied by partial condensation and rain, falling or suspended, called the *rain* stage. In the third, the temperature has fallen to 273°A. and the precipitated water freezes, while at the same time partial evaporation takes place, the temperature remaining constant, called the *hail* stage. In the fourth, all the water is frozen and the temperature still falls, or the *snow* stage is reached. These processes bring about different final results according as the condensed water is removed; and it must be remembered that in a reversed process, as that in which we seek to follow the changes in a mass of descending air, owing to the removal of the condensed vapor (in the form of rain, snow, or hail), the departures from initial conditions will be marked; in other words, the process is not, strictly, a reversible one. Neuhoff starts, not with the usual assumption of a kilogram of moist air, but with the assumption that, the condensed aqueous vapor being constant, then the weight of $1+x$ kilograms of

Method showing diminution of temperature with altitude

moist air during the ascent will be constant. The quantity x is the vapor that is mixed with the kilogram of dry air, and is designated as the "mixing ratio," while the quantity of aqueous vapor contained in 1 kilogram of moist air is the specific moisture. For high altitudes the mixing ratio is very small (from .001 to .003); that is to say, there are from 1 to 3 grams of aqueous vapor mixed with 1 kilogram of dry air. Referring to the adiabatic diagram (Fig. 57), dotted lines for each 5 grams indicate the constant quantity of moisture needed for saturation. The adiabats of the dry stage are the straight lines running parallel to the diagonals of the small squares, while those of the condensation stage are the curved lines indicated by dot and dash. An example of the use of the diagram is as follows: starting with a pressure of 1,012 kilobars, temperature 293° A. and relative humidity 80 per cent at the point A, the amount of vapor needed for saturation would be nearly 14 grams; but as the relative humidity is 80 per cent, the amount present is only 11.2 grams. The point of saturation and beginning of condensation will be found by following the adiabat for the dry stage until it intersects the gram line of saturation representing 11.2 grams at the point B. The conditions at B at the end of the dry stage are: temperature 288° A., pressure of vapor at saturation 17 kilobars; amount of water present 11.2 grams, pressure of atmosphere 960 kilobars; and altitude, 500 meters. The air is now saturated, and condensation begins. If the temperature falls to 283° A., the condensation adiabat is followed from the point B to its intersection with the isothermal of 283° A. at C. Here the conditions are found to be: pressure, 850 kilobars; vapor pressure of saturation, 11 kilobars; amount of water vapor, 9 grams. Therefore the quantity condensed is $11.2-9=2.2$ grams, the altitude 1,500 meters. When the temperature falls to 273° A. the rain stage ends. This is shown at D; and the corresponding pressure is 650 kilobars; vapor pressure, 6 kilobars. The quantity of vapor is 6 grams, showing that 5 grams have been condensed into rain. We now pass to an altitude of 3,650 meters, the third or hail stage, where there is a constant temperature until all the water is frozen, except the small

Mixing equations

quantity lost by evaporation. To freeze the 6 grams of water still left requires an ascent of about 30 meters per gram; and, therefore, the point E is about 180 meters higher, or about 3,830 meters; the pressure, 630 kilobars; the vapor pressure, 6 kilobars; the amount of ice, about 6 grams. With further cooling, the condensation is in the form of snow; the pressure, say, 500 kilobars, or half the original pressure; the temperature, 261° A.; the weight of vapor 3 grams and of frozen water 6 grams, leaving 2.5 grams as snow at an elevation of 5,200 meters. The diagram is faulty, however, in that the rain and snow may, and generally do, fall and separate from the ascending air.

Air that is saturated at an initial temperature of 303° A. at a pressure of 1,000 kilobars and ascends adiabatically will have a temperature of freezing, or 273° A., at an elevation of 7,360 meters. If the initial pressure be greater the height will be less, approximately 100 meters for each increase of 30 kilobars initial pressure.

An interesting application of the diagram is made by Neuhoff in the case of the foehn wind, and of course might also be made in connection with the other winds of this type, such as the chinook and the well-known Santa Ana, or "norther," of southern California. While in the case of ascending air the condensed vapor may or may not separate from the air column and move to another locality, in the case of descending air the presence or absence of the vapor is of the utmost importance in connection with the heat developed by compression. If the air has originally passed over a mountain ridge and lost much of its vapor load in the shape of rain or fog or low cloud, or even snow, then we can no longer pass backward over the adiabats. During the descent of the air no vapor is removed unless by wind action, and we can use only the dry-stage adiabat. Neuhoff gives the following example: At an initial temperature of 287° A. and a relative humidity of 60 per cent we find the saturation curve for 10 grams, and hence

$$10 \times 60 \div 100 = 6 \text{ grams for the mixing ratio.}$$

If, now, the air expands adiabatically, then, as shown by the intersection of the diagonal for the dry adiabat with

the 6-gram saturation line, the air will have a temperature of 278° A. at 900 meters' elevation. If further expansion takes place, the air follows the condensation adiabat and has a temperature of 273° A. at 1,750 meters, and the saturation value is then 4.6 grams. If the air rises to a mountain ridge of 3,000 meters, the temperature is 265° A. and the saturation quantity 3 grams. Now, at a point slightly over the summit, the precipitation is heaviest. Let us assume that the total quantity of moisture remaining is 3 grams.

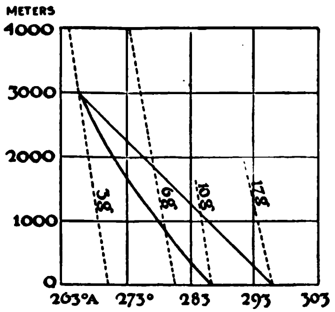


FIG. 58. FOEHN WIND

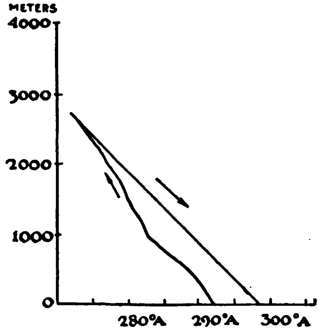


FIG. 59. FOEHN ADIABAT

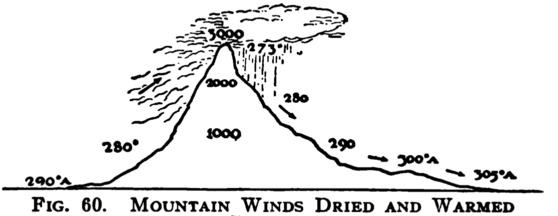


FIG. 60. MOUNTAIN WINDS DRIED AND WARMED IN DESCENDING

As the air descends on the other flank of the mountain there is adiabatic compression; and if we follow the dry adiabat, the temperature at the end — that is, at the initial height — would be 295° A. At this point it would require 17 grams to produce saturation; therefore the relative humidity is only 18 per cent, and the air is both warmer and drier. The above diagrams (Figs. 58–60), based upon diagrams by Neuhoff and Wegener, may be interesting.

For further remarks on the foehn, chinook, Santa Ana, and also the sirocco and hot "northers" of the western plains,

see the chapter on winds, where also are discussed the cold winds, the mistral of southern France, the bora of the Adriatic, the pampero of Argentina, and the southerly burster of Australia.

Mixtures

In several memoirs on the thermodynamics of the atmosphere, von Bezold treats of the changes which a given mass of air undergoes with elevation or horizontal convection from higher to lower pressure provided it is not affected by mixing with another mass of different temperature and moisture content. But mixing is constantly in operation, and it is therefore necessary to modify materially the equations in question.

James Hutton, of Edinburgh, was apparently the first to give attention to the problem of mixing in connection with the formation of rain. He was also the first to use the wet-bulb thermometer (used later by Leslie independently), and may have been led to his conclusions on the origin of rain from his studies of temperatures of evaporation. Hutton held that the mixing of two masses of air near saturation but of different temperature resulted in precipitation. For many years this view was accepted; then it was questioned, its opponents going so far as to say that no precipitation whatever could thus occur. As usual in such cases, the truth lies between the two extremes of opinion. Hann, in 1874, proved that, by mixture, condensation could indeed be produced, but that the method originally used for computing the amount of rainfall was inaccurate and that when proper corrections are applied the amount of precipitation is greatly diminished. In fact, the precipitation produced in this way can never be very great. Pernter, in 1882, computed various tables giving the possible quantities, and later von Bezold, reviewing the whole subject of the formation of precipitation, discussed the various causes, such as direct cooling, adiabatic expansion, and mixture. He showed that a mixture of saturated warmer air with unsaturated cooler air gives more condensation than a mixture of saturated cooler air with drier, warmer air. The flowing of a stream of saturated warm air into cool space is accompanied with more condensation than the reverse process. This explains why clouds of vapor

form so readily over an open warm-water surface while the formation of fog over cold surfaces is not so marked. Von Bezold illustrates the difference by calling attention to the fact that, although the opening of a wash-house door during moderately cool weather causes great clouds of vapor to pour out, the opening of an ice cellar on a warm day has not a similar result.

Conditions for maximum precipitation

To determine the maximum precipitation possible from mixing at different temperatures, von Bezold has constructed a set of twelve tables, of which an abstract follows, somewhat modified from the original:

Pressure 933 kilobars, difference of temperatures in 20-degree units:

First mass at temperature 273°A., saturation 100 per cent; second mass at temperature 253°A., saturation 100 per cent; maximum possible precipitation, 0.4 gram; final temperature of mixture, 264°A.

First mass at temperature 283°A.; second mass at 263°A., both at saturation; maximum precipitation, 0.5 gram; final temperature, 274°A.

First mass at temperature 293°A.; second mass at 273°A., both at saturation; maximum amount precipitation, 0.7 gram; final temperature, 284°A.

Pressure 933 kilobars, difference of temperatures in 10-degree units:

First mass at temperature 263°A., second mass at temperature 253°A., both at saturation; maximum precipitation, .04 gram; final temperature, 257°A.

First mass at temperature 273°A.; second mass at temperature 263°A.; saturation, 100 per cent; maximum precipitation, .1 gram; final temperature, 269°A.

First mass at temperature 283°A.; second mass at temperature 273°A.; saturation, 100 per cent; maximum precipitation, .2 gram; final temperature, 278°A.

First mass at temperature 293°A.; second mass at temperature 283°A.; saturation, 100 per cent; maximum precipitation, .2 gram; final temperature, 287°A.

Cooling more effective than mixing

These tables show how small is the precipitation obtained by mixture. A slight direct cooling will produce, therefore, as much precipitation as a considerable cooling due to mixture with cooler air even when saturated. The following is an illustration: If the first mass

has a temperature of 293° A., and the second of 273° A. (third case above), the maximum precipitation would be .7 gram and the final temperature 284° A., indicating a cooling of the whole volume 9°. By direct cooling of a single mass, the same amount of precipitation could be obtained by a fall from 293° to 292°; and by adiabatic expansion,— that is, cooling due to ascension, diminished pressure, and expansion at a temperature of 291°, which would occur when the mass was lifted about 300 meters. This example shows how slight need be the cooling (direct) by contact with cold surfaces, or through radiation, or even by rising, to produce condensation and precipitation equivalent to extensive mixing.

Super-saturation

Aqueous vapor can exist in a supersaturated state, a condition most likely to occur when the air is unusually pure and free from dust or nuclei on which condensation might occur. This will be referred to later in connection with the experiments of Aitken, Kiessling, Robert von Helmholtz, Wilson, Simpson, and others. Condensation and precipitation occur in such supersaturated masses of air when nuclei are present or when electric discharges take place.

42. Formation of fog at sea. In 1913 a special investigation of ice conditions, meteorology, and oceanography was undertaken by the Board of Trade of the British government for that part of the North Atlantic traversed by the large liners. The steamship *Scotia*, a whaler rigged as a three-masted bark with auxiliary steam, spent about fifteen weeks off the coasts of Newfoundland and Labrador. From the report of the meteorologist, Mr. G. I. Taylor, the following notes are taken regarding the temperatures and humidities recorded during the formation and dissipation of fog.[1] A captive-balloon ascent made on August 4, 1913, is particularly interesting; for it is possible to trace therefrom the effect on the temperature-height and humidity curves of two different changes,— one from cooling to warming, and the other from warming to cooling again. In Fig. 61 is shown the vessel's course and

Balloon record over the Atlantic

[1] This work has now been taken over by the U. S. Coast Guard vessels *Seneca* and *Tampa* (formerly the *Miami*).

position on preceding days. The center of a depression had passed over Belle Isle a few days before, the wind changing from one direction to another. The temperature changes were probably as follows: On July 29 the air blew from the land and was cooled by the water near the coast. Then it blew over warmer water until the morning of August 3, when the wind changed back toward the cold water. The effect of these changes is shown in Fig. 62. From sea level to 370 meters there is a negative temperature gradient (an increase in temperature with height, generally called an inversion)

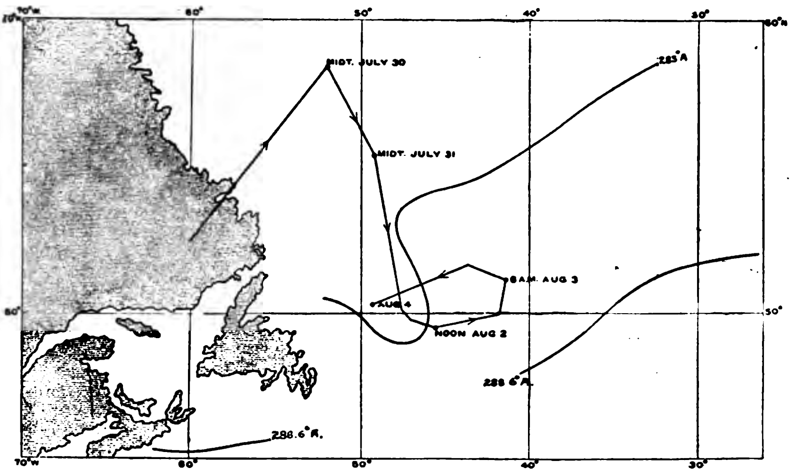


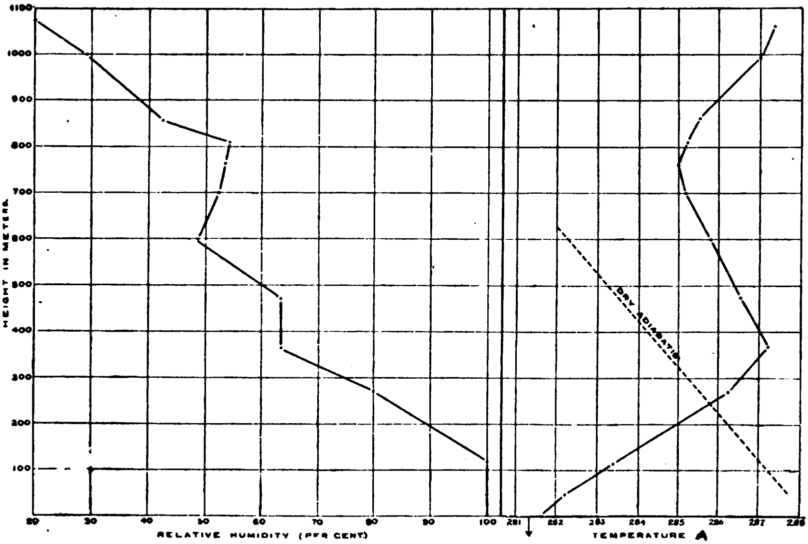
FIG. 61. SEA TEMPERATURE AND PATH OF AIR DURING FOG OFF NEWFOUNDLAND, S. S. SCOTIA, AUGUST 4, 1913

corresponding with the cooling from 8 A.M., August 3, until the time of the ascent. From 370 meters to 770 meters the temperature gradient is positive and corresponds with the warming which the air experienced from July 30 to August 3. Above 770 meters the gradient is negative again, and corresponds with the cooling as the air blew off the land on to the Arctic current.

The humidity curve indicates that the high temperatures in the highest layers explored were due to hot land rather than hot sea; the instrument registered the extremely low humidity of 20 per cent at the greatest height.

Measurements were made three times daily during May,

1915, aboard the ice-patrol cutter *Seneca*, of the number of persistent nuclei in the air per cubic centimeter by the corona method of Barus. Wells states that the number was



After Taylor

FIG. 62. SEA FOG TEMPERATURE. CONDITIONS DURING FOG, AUGUST 4, 1913, 7 P. M., S. S. SCOTIA

Weather, thick fog; wind direction, S. E. ½ S.; wind velocity, five miles per hour at all heights. The position of the arrow ( ) in the temperature-height diagram represents the temperature of the sea.

found to be never less than 400, normally 1,000, and on three occasions as high as 50,000. The nucleation was generally high in cyclonic areas, leading to the inference that the nuclei at sea are chiefly salt particles; i. e., evaporated spray. The amount of water in a cubic meter of fog was found, by evaporating the fog electrically and measuring the humidity at the higher temperature, to be 0.7 gram. The fog particles were found to have a diameter of the order of .0005 cm. A rise of 1.4° C. in temperature would dispel this fog; and therefore a slight temperature "inversion" resulted in a shallow fog, not extending as high as the masthead.

Marine fog

Von Bezold considers the following fogs and clouds as originating by mixture:

"The fog above warm, moist surfaces, under the influence of colder air, such as winter fogs at sea.

"The 'rank and file clouds,' occurring on the boundary of two strata flowing rapidly above each other; and which Von Helmholtz first recognized as a consequence of air motion and designated by the name 'atmospheric billows,' in which, however, adiabatic condensation also occurs at places where the air is thrown upward after the manner of the formation of crests and foam on ocean waves.

"The layers of stratus that also form at such separating surfaces. Cloud streamers that form and again dissolve at the summits of mountains or in narrow mountain passes when the topography is such as permits interpenetration of warm and cold air masses.

"The ragged clouds, or the disconnected clouds, such as occur during rapid motions of the air, perpetually changing their forms and appearing and disappearing."

It is because of these processes,—condensation, evaporation, compression, and expansion— that the motion of a cloud does not necessarily give a true measure of the motion of the air; for sometimes clouds hang apparently motionless on the mountains while strong winds stream through them (for example, the foehn cloud bank, the "Tablecloth" of Table Mountain, the cloud cap of the Helm wind, and others); and again, balloonists moving horizontally pass through long stretches of cloud masses. Within the cloud masses themselves marked circulation may occur. Observations by balloonists passing through large cumuli show that there are movements within the cloud that are independent of the general drift of the cloud as a whole, and that small bodies are whirled up and down, sometimes violently. In cumulo-nimbus, or thunder clouds, there is much turbulence irrespective of the progressive motion of the storm. It is difficult, however, to obtain reliable and sufficiently detailed records of air motion as well as the physical changes in temperature, humidity, and pressure. It may be pointed out that in so familiar a problem as recording temperature, our thermometers tell only a part of the story of the heat change; and our instruments for recording humidity are even less satisfactory than the thermometers. The ordinary mercurial thermometer indicates simply the difference in expansion

Turbulence within cloud

of a small quantity of mercury at the bulb, and of the glass. It may not even correctly give the temperature of the surrounding air; and of course it tells us nothing of the heat energy gained or lost in any change of state of the water vapor when air and vapor mix. The so-called "latent" heat (the name came into use at a time when heat was thought to be a material flow and is therefore misleading), the heat accompanying change from vapor to liquid and from liquid to solid, a change that is of constant occurrence in nature, is not indicated by any present form of thermometer. It may be well to repeat here that latent-heat energy may not always appear or reappear in the form of heat. Thus, as we shall see in the frost problem, it would be a mistake to regard the heat of condensation as actually set free and causative of a rise in temperature. It is the kind that may be utilized in work. The energy thus freed probably goes to reinforce the molecular energy expended by the body in cooling. It thus tends to retard the rate of cooling. It does act to prevent lowering of the temperature, but not as a quantity of heat directly available in raising the temperature.

Evaporation

43. The dissipation of aqueous vapor. We pass now from the consideration of problems dealing with the formation of fog, cloud, and rain to the reverse problem of dissipation of the vapor in either visible or invisible form. This brings us to the process of evaporation as contrasted with the reverse process of condensation.

If a mass of unsaturated air is supplied with water, there will be cooling, and the degree of cooling will vary with the dryness, being greater as the saturation deficit is greater; or, in other words, the greater the amount of water evaporated, the greater the degree of cooling.

Low temperature before dissipation

Von Bezold states from experience that frequently when passing through strata of fog such as fill the mountain valleys on calm, clear mornings, as one ascends the valley, the impression of colder weather occurs immediately on reaching the upper limit of the fog. It frequently happens that just before the sun dissipates the morning fog, a sensation of cold is experienced. He explains these phenomena on the assumption that the

temperature just below the upper boundary, when the fog is dissolving, is lower than that above or below. For when the sun begins to warm the upper side of the fog there occurs, first, relative dryness, which will, according to the rapidity of evaporation, extend somewhat into the fog layer. In other words, the layer of fog just under the upper surface has the lowest temperature; and above this the temperature rises and the humidity falls. This relation was strikingly brought out by Sigsfield, in a balloon trip over southern Germany, October 26, 1889. Gross, who also made a balloon voyage in the same year, found that in passing through thick clouds the temperature fell very low at the upper part of the cloud, but above this, rose decidedly.

In the cloud work at Blue Hill it was frequently noticed that the tops of cumuli were colder than was the air at a corresponding level at the same time.

In the matter of formation and dissolution a fog is but a cloud resting on earth, while a cloud is simply a raised fog.

Evaporation increase with warming

A steady increase in cooling or warming, if by radiation or expansion or compression, is accompanied by a steadily increasing evaporation or condensation; but in the case of mixture the process can continue and yet cause, first condensation, and later evaporation. The breath, exhaled into cool air, leaves the mouth saturated but not condensed; but on mixing with the air it is chilled, and we see the vapor in visible form. With further mixture with cool, dry air it dissolves.

Mixtures that cause dissolution

If we mix saturated cooler air with increasing amounts of warmer air, then the warming of the mixture proceeds more rapidly at first than subsequently, whereas in the reverse process cooling proceeds more slowly at first, and faster afterwards.

CHAPTER XIII

Dust and Microbes

44. Foreign matter in the atmosphere. Thus far we have dealt with the atmosphere as though it were made up only of gases and water vapor. It consists of something more than these. There is in suspension in the atmosphere a large amount of organic and inorganic matter. Moreover, these affect the processes of condensation and evaporation. They have not been studied much in connection with the latter physical process, but in condensation their influence is marked and has been exhaustively studied by Aitken, Barus, and others. The necessity of the presence of nuclei before condensation can occur, as pointed out by Aitken, will be referred to in a forthcoming paragraph.

The foreign matter, sometimes called the impurities of the air, is composed chiefly of microörganisms and dust. Bacteria abound in large numbers in the atmosphere of a city, and much work has been done by medical investigators to trace relationship between vitiated air and disease. For many years it was held that there was a direct relation between the two; but the later studies and experiments do not entirely confirm this view. We may not go into the discussion at length, but it may be of interest to quote here the conclusions of Hill and others, based on experiments at the physical laboratory of the London Hospital Medical College,[1] on the influence on health of the atmosphere in confined and crowded places. In brief these are:

Bacteria in the atmosphere

"No symptoms of discomfort, fatigue, or illness result from air rendered in the chemical sense impure by the presence of human beings, so long as the temperature and moisture are kept low. Such air can be borne for hours without any

[1] Recent experiments by Eastman and Lee referred to in *Science*, Aug. 11, 1916, p. 183, show that the harmfulness of respired air is not due to its chemical components. See also the researches of the New York State Commission on Ventilation, published in *American Journal of Public Health*, Vol. V (1915), p. 85. Little can be said at present regarding the effect of atmospheric conditions on

evidence of bodily or mental depression. . . . Heat stagnation is, therefore, the one and only cause of the discomfort, and all the symptoms arising in the so-called vitiated atmosphere of crowded rooms are dependent on heat stagnation. The moisture, stillness, and warmth of the air are responsible for all effects; and all the efforts of the ventilating engineer should be directed toward cooling the air in crowded places and cooling the bodies of the people by setting the air in motion. The essentials required of any good system of ventilation are (1) movement, coolness, proper degree of relative moisture of the air, and (2) reduction of the mass influence of pathogenic bacteria. The chemical purity of the air is of very minor importance."

Vitiated air and disease

Air laden with bacteria is generally considered offensive and conducive to the spread of disease. In certain hospital wards measurements of bacteria present have shown that fifty thousand recognizable organisms may exist in a cubic meter of air.

Dust content of the atmosphere

The principal sources of atmospheric dust are volcanic action, attrition of wind on land and sea, evaporation, and combustion. The movement of soil material by the winds is much greater than is generally supposed, and even in open places dust is being deposited and removed at a more rapid rate than in places where its presence is immediately perceptible. The movement of soil material by the wind is described in great detail by Free and Stuntz, in *Bulletin No. 68* of the Department of Agriculture, Bureau of Soils. Students of geology know of the extensive influence of dust in such problems as dune control, desert geology, the occurrence of the loess, and volcanic dust. Black, at Edinburgh, in 1902 showed that the amount of dust deposited in an open rain gauge having a funnel 150 millimeters in diameter varied from 1.62 grams to 10.37 grams per month. The average was 2.7 grams, which is equivalent to 150 grams per square meter per month, or

the nervous system, nor do we know much about the relation between metabolic phenomena and atmospheric conditions.

Huntington in his book on *Civilization and Climate* has shown that the maximum physical efficiency occurs at intermediate seasons, and that the optimum temperature of the outside air for the physical work of human beings is 288°A. and for mental work 277°A.

1.8 kilograms per square meter per year (5,157 tons per square mile). Fry, at Cincinnati, in work done for the Smoke Abatement League, collected dust in buckets kept partly filled with water and placed on the roofs of various buildings. The dust was filtered off, extracted with concentrated hydrochloric acid, and weighed. The average monthly amount was 60.2 grams per square meter, or 723 grams per square meter for the year (2,062 tons per square mile). The material collected was carbonaceous and ashy, probably derived largely from coal smoke.

Enormous quantities estimated

Between March 9 and 12, 1901, dust fell at various places in Europe, and an effort was made by Hellmann and Meinardus to estimate the quantity. The amounts varied from 11.23 grams to 1 gram per square meter (31 to 3 tons per square mile). The values were largest in southern Europe, decreasing toward the north. The area covered was at least 300,000 square miles of land surface and 170,000 square miles of ocean. The estimates are somewhat doubtful, but it is plain that the total quantity of sirocco dust which falls in Europe is very large. The authors are of the opinion that if a storm as violent as that of March, 1901, occurred once in five years, there would have been 143 centimeters of desert material carried into Europe in three thousand years.

African dust in southern Europe

Measurements of the monthly deposit of soot in Pittsburgh, during the year ending March, 1913, showed that at the point of maximum deposit the annual rate was nearly 4 kilograms per square meter (1,950 tons per square mile), and at the place of minimum deposit nearly 1 kilogram per square meter (600 tons per square mile). Measurements made at Leeds, England, serve as a basis for estimating that there are annually introduced into the air near that city 35,000,000 kilograms of soot.

Pittsburgh's annual dust deposit

John Aitken of Edinburgh has probably done more than any one else in studying the effect of dust as nuclei of cloudy condensation. His several papers in the *Transactions* of the Royal Society of Edinburgh show the mean limit of visibility for different winds and the relation to the dust content. By means of a dust counter he has shown that the product

of the number of particles per unit space by the distance or limit of visibility is constant for equal depressions of the wet bulb. It appears that for a depression of 3 degrees, 1,000 particles per cubic centimeter would completely obscure large objects at a distance of 100 miles; that 100 times the number of particles would completely obscure objects at a distance of 1 mile; and a million particles would obscure objects at one tenth of a mile. In large cities the number of particles may exceed 300,000 per cubic centimeter even in fine weather.

Obscuration of objects

Carl Barus, in his paper on "Condensation of Atmospheric Vapor,"[1] shows how cloudy condensation depends on air temperature and dust content. Aitken has shown that condensation takes place only upon free surfaces when saturation temperature exists and only when sufficient nuclei are present. His experiments suggested a possible method of counting dust particles which, for the most part, are too small to be seen even with a microscope. By making the extremely small particles as well as the larger ones centers of condensation, that is, making them the nuclei of small raindrops, it is practicable to count the drops and determine the number of particles. By mixing a small quantity of dusty air with a large quantity of dustless air, and allowing the particles to fall on a micrometer, they can be counted by the aid of a magnifying glass. Then, knowing the proportion of dustless to dusty air and allowing for dilution, the number of particles can be estimated. The general plan of Aitken's dust counter is shown in the diagram (Fig. 63). *A* is the test receiver where the air under investigation is introduced and the particles are counted. It is an ordinary glass flask with flat bottom, supported in an inverted position. *B* is an air pump connected with *A* by an India-rubber tube *C*. The pump *B* is drawn in the position shown in the diagram for convenience of illustration; in practice, it is placed above the level of the table for convenience in using the pump while the eye is

Precondition of condensation

Dust counter

[1] *Weather Bureau Bulletin No. 12*, 1893, and *Smithsonian Contributions*, Vol. XXXIV, 1905. See also *The Nucleation of the Uncontaminated Atmosphere* by Barus and Pierce, Carnegie Institution, Jan., 1906.

watching at the magnifying glass. *D* is a cotton-wool filter connected with *A* by means of the pipe *E*. The pipes *C* and

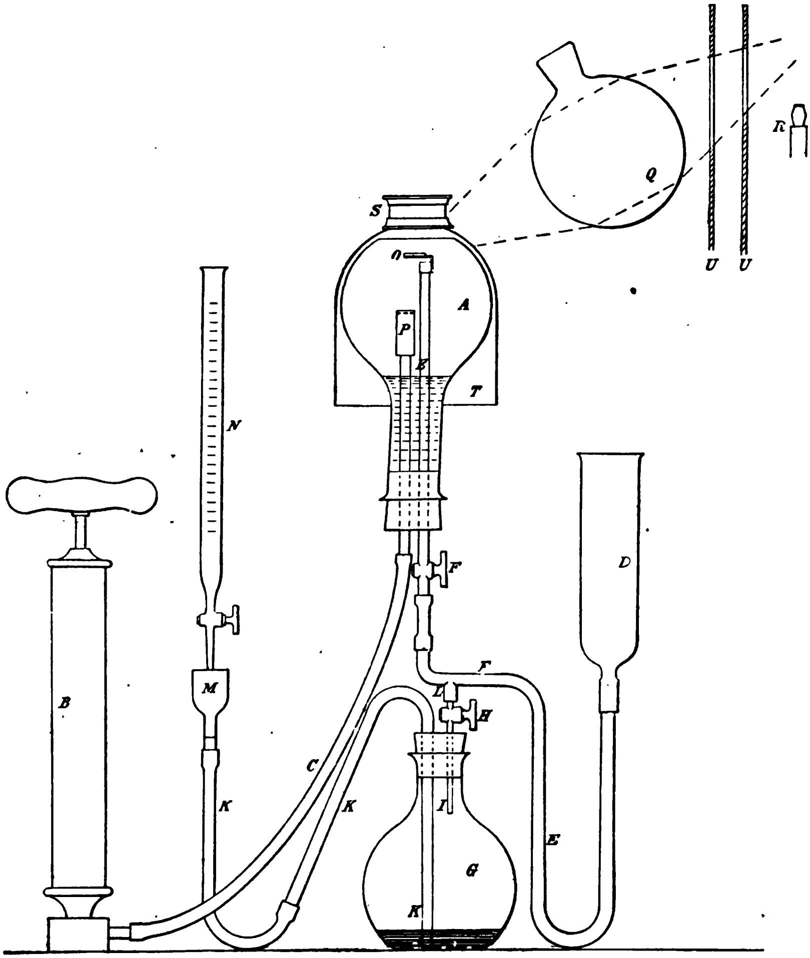


FIG. 63. THE ORIGINAL DUST COUNTER (AITKEN)

E pass through an India-rubber stopper in *A* and project upward into the receiver; *C* stops about the middle, while *E* rises to near the top and forms the support to which the counting stage *O* is attached. *F* is a stopcock for closing the connection between the receiver *A* and the filter *D*.

To keep the air under examination saturated with vapor, some water is put in the receiver A, and from time to time the receiver is inverted. The stage O, on which the drops are counted, is a small plate of highly polished silver about 1 centimeter square, ruled with fine lines at right angles to each other and 1 millimeter apart. The stage is supported exactly 1 centimeter below the flat top of the receiver. Otherwise the stage would become covered with a heavy deposit of dew, when it is of no service. When mounted as shown the dew is easily cleared away by heating the tube E. The heat is carried forward to the stage by the entering air. The stage is not placed centrally over the pipe E, because if the stage is too hot the drops roll away and quickly evaporate; on the other hand, if they are too cold the surface becomes wet and counting is impossible. The stage is viewed through the bottom of the flask, using the magnifying glass S. The following example shows the manner of estimating the number of particles in a sample of air.

In addition to the number of drops per square millimeter, the quantities required to complete the estimate are the capacities of A and B. If A has a capacity of 500 cubic centimeters and there are 50 cubic centimeters of water, its air capacity is reduced to 450 cubic centimeters. If into this pure air we introduce 1 cubic centimeter of the air to be tested, the dusty air will be, so to speak, diluted 450 times. But the air is not only diluted; it is also expanded by the pump, which has a capacity of 150 cubic centimeters. The dust that was in the original 1 cubic centimeter is thus expanded by the two processes into 600 cubic centimeters. The number of drops per cubic centimeter counted on the stage must, therefore, be multiplied by 600 to give the number in the original cubic centimeter of dusty air. Suppose that we counted one drop per square millimeter, then as there is 1 centimeter of air above the stage, this will give 100 drops per cubic centimeter in the diluted and expanded air; and this, multiplied by 600, gives 60,000 dust particles per cubic centimeter of the air tested.

Aitken describes the many precautions necessary in using the apparatus and the difficulties he met in his investigations.

He points out, too, that a very slight degree of supersaturation will cause condensation of some of the dust particles in the air; but that the degree of supersaturation which is sufficient to cause some of the particles to become active centers is yet insufficient to cause condensation to take place on others. This might indicate that the condensing power of dust particles is affected by their size. Some of the values obtained by Aitken are:

NUMBER OF DUST PARTICLES IN AIR

Source	Number per c.c.
Outside (raining)	32,000
Outside (fair)	130,000
Room	1,860,000
Room (near ceiling)	5,420,000
Bunsen flame	30,000,000

These numbers are very far from being constant, and vary with conditions. In one of his papers[1] Aitken mentions that a cigarette smoker sends 4,000,000,000 particles, or more, into the air with every puff. About the smallest number of particles observed is 500 per cubic centimeter. The following are some of the conclusions arrived at:

The earth's atmosphere is greatly polluted with dust produced by human agency. This dust is carried to a considerable elevation by the hot air rising over cities.

The transparency of the air depends on the number of dust particles in it and also on its humidity. The less dust, the more transparent the air; and the dryer the air the more transparent it is. There is no evidence that humidity alone, that is, water in its gaseous state and apart from dust, has any effect on the transparency.

The dust particles in the atmosphere have vapor condensed on them though the air may not be saturated.

The amount of vapor condensed on the dust in unsaturated air depends on the relative humidity and also on the absolute humidity. The higher the humidity and the higher the vapor tension, the greater is the amount of moisture held by the dust particles when the air is not saturated.

Haze is generally produced by dust; and if the air be dry the vapor has little effect and the density of the haze depends chiefly on the number of particles present.

[1] Royal Soc. of Edin., *Proceedings*, Feb. 4, 1889.

Barus and Pierce have made many measurements of atmospheric nucleation. The details of these experiments cannot be given here; but the general deductions are of much interest, such as the extremely high nucleation found in winter months as compared with summer months; again the efficiency of rain in depressing nucleation; and finally the totally different character of the curves in different years. Simultaneous measurements made at Providence and at Block Island show that the nucleations at the former station are much in excess, perhaps as much as fifty times greater. The difference may be due, since it is exaggerated in the winter months, to the originally ionized products of combustion.

Nucleation of unfiltered air

Aitken devised what he called a "koniscope," in which the color of transmitted light is determined by the size of the cloud particles, the depth of color indicating the number of particles present. The indications of the koniscope have been compared with the number given by the dust counter as follows:

The koniscope

Dust counter particles per c.c.	Koniscope, depth of color
50,000..............	Color just visible
80,000..............	Very pale blue
500,000..............	Pale blue
1,500,000..............	Fine blue
2,500,000..............	Deep blue
4,000,000..............	Very deep blue

There is great diversity in the action of different kinds of dust particles in producing fogs. In ordinary country fogs the dust particles are similar to those in clouds. If the condensation be made quickly, then a process of differentiation takes place, the smaller particles evaporating and the larger ones increasing in size. In this way clouds and ordinary fogs tend to rain themselves out of existence. Not so a town fog in which many of the particles forming the nuclei of condensation have an affinity for water. Such affinity is fatal to differentiation of the particles, for by checking the same it prevents the natural decay and

Diversity in action of cloud particles

Dust with affinity for water

falling of the water particles. Air with dust particles having an affinity for water tends to produce a maximum number of small water particles with but little tendency to fall; whereas, if there be no affinity, the tendency is to produce a minimum number of large particles with a tendency to fall. The one kind of dust particle forms a persisting fog, while the other forms a fog with a tendency to rain itself away. This difference seems to account for the greater thickness and persistence of town fogs. There is considerable sulphur dioxide in city air, probably resulting from the burning of coal; and this serves to increase the formation and maintenance of fog. Aitken, in his paper on the sun as a fog producer, shows that under the influence of sunshine nuclei are formed which have such an affinity for water that condensation sets in at temperatures above the saturation temperature. Sulphur dioxide, while kept in pure air, shows little tendency to produce nuclei, but combines readily with other products of combustion and then, as Aitken puts it, "falls from its high state of a free-moving gaseous molecule to the condition of a solid or liquid particle confined to Brownian movements; and probably finds its independent existence in a fog particle or possibly in a raindrop."

Why town fogs persist

It must be pointed out that, while there may be several million dust particles in a cubic centimeter of air, small as these are they are not to be confused with gaseous particles. The number of molecules in a cubic centimeter of a gas under standard conditions is 2.7 billion billion, or, as it is generally expressed, 2.704×10^{18}. To get a better idea of the size of atoms and molecules it may be said that it requires 11,200 cubic centimeters of hydrogen to weigh one gram, and in that gram there would be over 300 thousand billion billion molecules and twice as many atoms, that is, 6.05×10^{23}. The energy carried by a beam of sunlight varies from day to day with the clearness of the air and the altitude of the sun. Atmospheric transmission is much dimished by perceptible haziness. Even in clearest weather and at high mountain stations there is an appreciable loss of energy in the passage through the air, and the percentage of

Gaseous and dust particles distinct

loss increases steadily as the wave length of the radiation becomes smaller. At Mount Wilson, under favorable conditions, almost 99 per cent of the infra-red radiation—the long waves—gets through; in the green part of the spectrum about 90 per cent reaches the earth; in the violet, 80 per cent; and in the ultra-violet, about 60 per cent. In certain parts of the infra-red the atmosphere is almost impervious to solar radiation. Aside from the actual absorption of the energy by the water vapor, there is a scattering of the radiation in directions other than that of the direct ray. If this did not occur, the sky on a clear day would appear black except for whatever whitish dust haze might be present. Lord Rayleigh has shown that the blue color of the sky is the result of this irregular dispersion, or scattering, of light, even in dust-free air, by the gas molecules. The molecules act with greatest effect on the short wave length, and therefore the scattering power for the violet and ultra-violet waves is ten times greater than for red light; hence the larger diffusion of blue. Dust particles, on the other hand, reflect light of all colors equally, and when the air is full of dust we have a mixed white and blue sky, or, as it is called, milky white. When the sun is near the horizon the beam has a much longer column of air to traverse, and the blue waves are less effective; then the sun looks orange or red. Furthermore, when the air is full of fine volcanic dust, such as followed the eruption of Krakatau, August 27, 1883; or of Katmai, June 6, 1912; or of La Soufrière and Pelée in 1902; or of Taal, January 27, 1911, then there is a marked absorption, from five to ten times the normal amount, and a marked coloring of the sky at sunset and sunrise. In the eruption of Krakatau large quantities of dust were thrown high in the air over the Sunda Strait, and this was carried slowly westward around the world, causing noticeably red sunsets for a period of more than twenty months in temperate latitudes. There were also halo effects and the so-called Bishop's ring (named after the observer at Honolulu). These last were due to diffraction of the light

Dust and light transmission

Why the sky is blue

Why sunsets are red

Effect of volcanic dust

in passing through the dust. Abbot, Fowle, Kimball, and others have studied these relations.

One source of atmospheric dust is the spray blown inland from the seas, containing the sodium chloride known to be present in certain quantity in rain. The amount of sodium chloride decreases with distance from the seashore; Du Bois gives yearly averages varying from 0.66 to 30 milligrams per liter of rain. He calculates the annual amount deposited on the dunes of Holland to be at least 6 million kilograms (13,227,720 pounds). The mean proportion of sodium chloride in rain in England is 2.2 milligrams per liter; at Rothamsted it is 2.01 milligrams per liter; at Nantes, France, it is 14 milligrams; and at Troy, New York, 2.7 milligrams.

Salt content of the atmosphere

OPTICAL PHENOMENA

45. Halos and coronas. Halos are effects caused by the refraction and reflection of the rays of the sun or moon by ice crystals. There are many different types of halo, no fewer than fourteen having been enumerated by Hastings. The commonest form is the halo of 22°. There is a close relation between this halo and the occurrence of cirro-stratus cloud. It is a luminous ring with the inner edge sharply defined and showing red. Proceeding outward, one may detect orange, yellow, and sometimes green, but seldom if ever violet. The sun (or moon) is the center of the circle; and usually one sees only the whitish ring with an inner edge of brownish red. Halos may last for several hours, according to the duration and thickness of the cirro-stratus cloud sheet. They are good indicators of coming storm conditions, for the reason that they depend upon the prevalence and intensity of the upper cloud layer. The frequency of halos and subsequent precipitation has been studied for Blue Hill Observatory by Palmer.[1] A detailed description of the different forms of halos by Besson was published in 1911.[2]

Colors of the halo

Halos and precipitation

[1] *Monthly Weather Review*, July, 1914.

[2] A translation is given in the *Monthly Weather Review*, July, 1914, p. 436. See also same author on the Halos of Nov. 1–2, 1913, p. 431.

Coronas are diffraction phenomena, frequently seen about the moon closely surrounding the source of light. They are seldom more than two degrees in radius; and when showing prismatic colors the red is at the outer edge. Solar coronas due to passing clouds must not be confused with the solar corona proper.

Corona

Parhelia (mock suns) are luminous spots about 22 degrees distant from the sun. Paraselenae (mock moons) are similar whitish spots produced by clouds passing before the moon. They have the usual red coloring on the edge nearest the source of light. When near the horizon they are elongated vertically, and thus distorted they may be mistaken for fragments of a rainbow. Parhelia disappear when the solar altitude exceeds 51 degrees. There is a well-known type of halo with a radius of 46 degrees. When a cloud that has caused a parhelion passes to 46 degrees above the sun a circumzenithal arc is formed.

Parhelia

Paraselenae

The anthelion (counter sun) is a round, luminous spot 180 degrees from the sun. It must not be confounded with the antisolar corona, or "glory" of the aëronaut.

Anthelion

A light pillar is a train of light extending vertically above the sun or moon; and it may also be prolonged beneath the luminary. It must not be mistaken for the luminous rays that diverge in all directions when sunlight is streaming through breaks in the clouds. The light pillar is always vertical.

It is perhaps proper to remark that a halo described by Bravais as the most authentic of all extraordinary halos,—the 90-degree halo of Helvelius (reported as seen in 1661),—does not really exist, according to investigations made by Hastings. This writer insists, moreover, that in explaining the various halo forms only such forms of ice crystals as are known to exist can be taken account of; further, that the orientation of the falling crystals must conform to the law of mechanics; and, finally, that all those features of halos attributable to reflections must find their explanation in every case in total reflections. Much,

Faulty data regarding halos

therefore, that has been advanced regarding halos is still unproved.

In foggy weather an observer, especially if he is on a height standing with his back to the sun, will see the shadow of his head or body cast upon the fog, surrounded by a colored ring of light, variously called "glory," "Ulloa's ring," "Bröcken specter." The green and red patches occasionally seen in cirrus clouds are known as "irisation."

"Glory," "Irisation"

The colors of rainbows, as well as the extent and position of greatest luminosity, depend upon the size of the drops producing the bow. It has been erroneously assumed that all rainbows show the same sequence of colors and have the same radius. If the sequence of colors in the primary bow commencing with the red is noticed, also the color showing maximum luminosity and which color band is widest, the size of the drop can be determined. Light is both reflected and refracted at the surface boundary of air and water; and a rainbow is, therefore, the result of reflection, refraction, and interference combined. Since each illuminated point on a drop of water in midair receives light from the whole disk of the sun, there are many overlapping spectra. The width and degree of separation of the interference bands is increased as the size of the drops diminishes. According to Pernter, rainbows of richest color are produced by drops varying from 0.2 to 0.4 millimeter in diameter.

Rainbows

CHAPTER XIV

Atmospheric Electricity

46. Thunderstorm phenomena. There are few more pronounced manifestations of atmospheric motion than a well-developed thunderstorm. While the lightning discharges are perhaps the most spectacular feature, there is much more to a storm of this character than the electrical display; and, indeed, it is doubtful if even what we see of the lightning is more than a small part of the phenomena and the conversion and dissipation of energy.

Franklin's experiment

The identification of lightning with electricity was the work of Franklin, and the classical kite experiment is described by him in a letter dated October 19, 1752 (old style). Very few investigators have repeated the experiment as described by Franklin, and, as a matter of fact, the results are somewhat different from those he described.[1] He did not draw the lightning from the clouds, as is so generally stated, but did obtain moderate induction effects; and these can be obtained without much danger on the approach of heavily charged clouds; but any direct discharge or flash of lightning will demolish both kite and string and probably injure the observer. Thus at Blue Hill Observatory during a kite flight, March 6, 1913, while there had been the usual static discharges, there was no lightning and no thunder previous to a discharge at 12:41 P.M., when 1,500 meters of steel wire were volatilized, and the observers stunned.

[1] Rotch, in *Science*, Dec. 14, 1906, shows that Franklin did not fly his kite until later in summer than June, 1752, or some two months later than has generally been supposed. Franklin had already prepared for publication precise directions for placing lightning rods upon all kinds of buildings. The letter to Collinson is dated Oct. 19, 1752. It reads:

"Make a small cross of light sticks of cedar, the arms so long as to reach to the four corners of a large thin silk handkerchief when extended. Tie the corners of the handkerchief to the extremities of the cross, so you have the body of a kite, which being properly accommodated with a tail, loop and string, will rise in the air like those made of paper, but being made of silk is better fitted to bear the wet and wind of a thunder gust without tearing. To the

An excellent résumé of recent investigation of thunderstorm phenomena is given by Humphreys in the *Monthly Weather Review*, June, 1914, from which the following extract is taken:

"A thunderstorm is a storm characterized by thunder and lightning, just as a dust storm is one characterized by a great quantity of flying dust. But the dust is never in any sense the cause of the storm that carries it along, nor, so far as known, does either thunder or lightning have any influence on the course—genesis, development, or termination—of even those storms of which they form, in some respects, the most important features. No matter how impressive or how terrifying these phenomena may be, they never are anything more than mere incidents to or products of the peculiar storms they accompany, as will be made clear by what follows. In short, they are never in any sense either storm-originating or storm-controlling factors.

Origin of atmospheric electricity

"A knowledge, or at least a good working hypothesis, of how the great amount of electricity incident to thunderstorms is generated, is absolutely essential to their logical explanation; that is, to a clear understanding of the probable interrelations between their many phenomena. Fortunately such an hypothesis, or theory rather, since it is abundantly supported by observations and by laboratory experiments, is available as a result of work done on this subject in India by Dr. G. C. Simpson[1] of the Indian Meteorological Department.

"Dr. Simpson's observations were obtained at Simla, India, at an elevation of about 7,000 feet above sea level, and covered all of the monsoon seasons, that is, roughly,

top of the upright stick of the cross is to be fixed a sharp pointed wire rising a foot or more above the wood. To the end of the twine next the hand is to be tied a silk ribbon and where the silk and twine join a key may be fastened. This kite is to be raised when a thunder gust appears to be coming on, and the person who holds the string must stand within a door or window or under some cover so that the silk ribbon may not be wet: and care must be taken that the twine does not touch the frame of the door or window. As soon as the thunder clouds come over the kite the pointed wire will draw the electric fire from them and the kite with all the twine will be electrified, and stand out every way and be attracted by an approaching finger. And when the rain has wet the kite and twine you will find the electric fire stream out plentifully from the key on the approach of your knuckle."

[1] *Memoirs, Indian Met. Dept.*, Simla, Vol. XX (1910), Pt. 8.

April 15 to September 15, of 1908 and 1909. He also obtained observations of the electrical conditions of the snow at Simla during the winter of 1908–1909.

"A tipping-bucket rain gauge gave an automatic continuous record of the rate and time of rainfall, while a Benndorf self-registering electrometer marked the sign and potential of the charge acquired during each two-minute interval. A second Benndorf electrometer registered the potential gradient near the earth, and a coherer of the type used in radio telegraphy registered the occurrence of each lightning discharge.

"All obvious sources of error were examined and carefully guarded against. Hence it would seem that the conclusions drawn from the thousands of observations given in the memoir are fully justified; and especially so since several independent series of similar observations made at different times, by different people and at places widely separated, have given confirmatory results in every case. Simpson's records show that—

Simpson's records

"(1) The electricity brought down by the rain was sometimes positive and sometimes negative.

"(2) The total quantity of positive electricity brought down by the rain was 3.2 times greater than the total quantity of negative electricity.

"(3) The period during which positively charged rain fell was 2.5 times longer than the period during which negatively charged rain fell.

"(4) Treating charged rain as equivalent to a vertical current of electricity, the current densities were generally smaller than 4×10^{-15} amperes per square centimeter; but on a few occasions greater current densities, both positive and negative, were recorded.

"(5) Negative currents occurred less frequently than positive currents, and the greater the current density the greater the preponderance of the positive currents.

"(6) The charge carried by the rain was generally less than 6 electrostatic units per cubic centimeter of water, but larger charges were occasionally recorded, and in one exceptional storm (May 13, 1908) the negative charge exceeded 19 electrostatic units per cubic centimeter.

"(7) As stated in paragraph (3) above, positive electricity was recorded more frequently than negative, but the excess was the less marked the higher the charge on the rain.

"(8) With all rates of rainfall positively charged rain occurred more frequently than negatively charged rain, and the relative frequency of

positively charged rain increased rapidly with increased rate of rainfall. With rainfall of less than about 1 millimeter in two minutes, positively charged rain occurred twice as often as negatively charged rain, while with greater intensities it occurred 14 times as often.

"(9) When the rain was falling at a less rate than about 0.6 millimeter in two minutes, the charge per cubic centimeter of water decreased as the intensity of the rain increased.

"(10) With rainfall of greater intensity than about 0.6 millimeter in two minutes the positive charge carried per cubic centimeter of water was independent of the rate of rainfall, while the negative charge carried decreased as the rate of rainfall increased.

"(11) During periods of rainfall the potential gradient was more often negative than positive, but there were no clear indications of a relationship between the sign of the charge on the rain and the sign of the potential gradient.

"(12) The data do not suggest that the negative electricity occurs more frequently during any particular period of a storm than during any other.

"Concerning his observation on the electrification of snow Dr. Simpson says:

"'As far as can be judged from the few measurements made during the winter of 1908–1909 it would appear that:

"'(1) More positive than negative electricity is brought down by snow in the proportion of about 3.6 to 1.

"'(2) Positively charged snow falls more often than negatively charged.

"'(3) The vertical electric currents during snowstorms are on the average larger than during rainfall.

"'(4) The charge per unit mass of precipitation is larger during snowfall than during rainfall.'

"While these observations were being secured, a number of well-devised experiments were made to determine the electrical effects of each obvious process that takes place in the thunderstorm.

"Freezing and thawing, air friction, and other things were tried, but none produced any electrification. Finally, on allowing drops of *distilled* water to fall through a vertical blast of air of sufficient strength to produce some spray, positive and important results were found, showing:

"(1) That breaking of drops of water is accompanied by the production of both positive and negative ions.

"(2) That three times as many negative ions as positive ions are released.

"Now, a strong upward current of air is one of the most conspicuous features of the thunderstorm. It is always evident in the turbulent cauliflower heads of the cumulus cloud, the parent, presumably, of all thunderstorms. Besides, its inference is compelled by the occurrence of hail, a frequent thunderstorm phenomenon, whose formation requires the carrying of raindrops and the growing hailstones repeatedly to cold and therefore high altitudes. And from the existence of hail it is further inferred that an updraft of at least 8 meters per second must often occur within the body of the storm, since, as experiment shows, it requires approximately this velocity to support the larger drops, and even a greater velocity to support the average hailstone.

"Experiment also shows that rain cannot fall through air of ordinary density whose upward velocity is greater than about 8 meters per second, or itself fall with greater velocity through still air; that in such a current, or with such a velocity, drops large enough, if kept intact, to force their way down, or, through the action of gravity, to attain a greater velocity than 8 meters per second with reference to the air, whether still or in motion, are so blown to pieces that the increased ratio of supporting area to total mass causes the resulting spray to be carried aloft or left behind, together with, of course, all original smaller drops. Clearly, then, the updrafts within a cumulus cloud frequently must break up at about the same level innumerable drops which, through coalescence have grown beyond the critical size, and thereby, according to Simpson's experiments, produce electrical separation within the cloud itself. Obviously, under the turmoil of a thunderstorm, its choppy surges and pulses, such drops may be forced through the cycle of union (facilitated by any charges they may carry) and division, of coalescence and disruption, from one to many times, with the formation on each at every disruption, again *according to experiment* of a correspondingly increased electrical charge. The turmoil compels mechanical contact between the drops, whereupon the charges break down the surface tension and insure coalescence. Hence, once started, the electricity of a thunderstorm rapidly grows to a considerable maximum.

Breaking-drop theory

"After a time the larger drops reach, here and there, places below which the updraft is small—the air cannot be rushing up everywhere—and then fall as positively charged rain, because of the processes just explained. The negative electrons in the meantime are carried up into the higher portions of the cumulus, where they unite with the cloud particles and thereby facilitate their coalescence into negatively charged drops. Hence, the heavy rain of a thunderstorm should be positively charged, as it almost always is, and the gentler portions negatively charged, which very frequently is the case.

"Such, in brief, is Dr. Simpson's theory of the origin of the electricity in thunderstorms, a theory that fully accounts for the facts of observation and in turn is itself abundantly supported by laboratory tests and simulative experiments.

"If this theory is correct, and it seems well founded, it must follow that the one essential to the formation of the giant cumulus cloud, namely, the rapid uprush of moist air, is also the one essential to the generation of the electricity of thunderstorms. Hence the reason why lightning seldom, if ever, occurs except in connection with a cumulus cloud is understandable and obvious. It is simply because the only process that can produce the one is also the process that is necessary and sufficient for the production of the other.

"*The violent motions of cumulus clouds.* From observations, and from the graphic descriptions of the few balloonists who have experienced the trying ordeal of passing through the heart of a thunderstorm, it is known that there is violent vertical motion and much turbulence in the middle of a large cumulus cloud, a fact which, so far as it relates to the theory alone of the thunderstorm, it would be sufficient to accept without inquiring into its cause. However, to render the discussion more nearly complete, it perhaps is worth while, since it is a moot question, to inquire what the probable cause of the violent motions in large cumulus clouds really is—motions which, in the magnitude of their vertical components and degree of turmoil, are never exhibited by clouds of any other kind nor met with elsewhere by either manned, sounding, or pilot balloons.[1]"

[1] Simpson's views are given at length in a paper on "The Electricity of Atmospheric Precipitation," *Phil. Mag.*, S. 6, Vol. XXX (1915), p. 1.

47. Cause of the turbulence in thunderstorms. Humphreys thinks that the difference between the actual temperature gradient of the surrounding atmosphere and the gradient (adiabatic) for saturated air within the cloud itself is the cause of the turbulence. Assume the temperature to be 303°A. and the dew point 288°A. The adiabatic decrease, as we have seen, is approximately one degree per hundred meters; therefore condensation would begin at a height of 1.5 kilometers. It may be remarked, however, that we have no data justifying the use of this rate at times of thunderstorms. The cloud particles are carried up, and at an elevation of approximately half a kilometer above the plane of condensation the drops lag or drop. Hence for two reasons,—because the heat of the condensed water is no longer available to the air from which it was condensed, and because little heat is available from further condensation,— the rate of decrease again approaches the assumed adiabatic gradient for dry air.

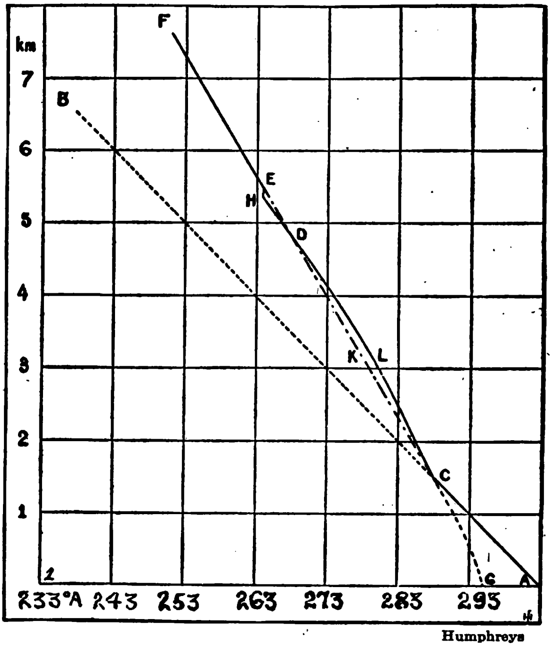


FIG. 64. TEMPERATURE GRADIENTS WITHIN (CLD) AND WITHOUT (CKD) CUMULUS CLOUDS

In the accompanying diagram (Fig. 64) *AB* is the approximate temperature gradient for nonsaturated air. *GCKDEF* is the supposed temperature gradient before convection begins; that is, the average rate as deduced from many soundings, or about 0.6 degree instead of 0.96, except near the surface where the gradient is even less and sometimes negative.

To explain the abrupt temperature change Humphreys says: "As convection sets in, the temperature decrease near the surface soon approximates the adiabatic gradient for dry air, and this condition extends gradually to greater altitudes, till, in the assumed case, condensation begins at the level *C*, or where the temperature is 283°A. Here the temperature decrease, under the assumed conditions, suddenly changes from 10 degrees per kilometer increase of elevation to rather less than half that amount, but slowly increases with increase of altitude and consequent decrease of temperature. At some level, as *L*, the temperature difference between the rising and the adjacent air is a maximum. At *D* the temperature of the rising air is the same as that of the air adjacent, but its momentum presumably carries it on to some such level as *H*. Within the rising column, then, the temperature gradient is approximately given by *ACLDHE*, and that of the surrounding air by *ACKDEF*.

"The cause, therefore, of the violent uprush and turbulent condition within large cumulus clouds is, presumably, the difference between the temperature of the inner or warmer portions of the cloud itself and that of the surrounding atmosphere at the same level, as indicated by their respective temperature gradients *CLD* and *CKD*. Clearly, too, while some air must flow into the condensation column all along its length, the greatest pressure difference, and, therefore, the greatest inflow, obviously is at its base. After the rain has set in, however, this basal inflow is from immediately in front of the storm, and necessarily so, as will be explained later."

Difference of temperature in cloud

Since rapid vertical convection of humid air is essential to the production of the thundercloud, the conditions under which the vertical temperature gradient necessary to this convection can be established must be (1) strong surface heating, especially in regions of light wind; (2) the overrunning of one layer of air by another at a temperature sufficiently low to induce convection; and (3) the underrunning and consequent uplift of a saturated layer of air by a denser layer.

The turbulent character of the air motion and the irregularity of cloud mass are shown in Figs. 65 and 66.

FIG. 65. TURBULENCE IN THUNDERSTORM CLOUDS McAdie

FIG. 66. TURBULENCE IN THUNDERSTORM CLOUDS McAdie
Fig. 66 was taken four minutes later than Fig. 65.

48. Conditions favorable to thunderstorms. The more humid the air and the more energetic the local convections the greater the frequency and intensity of thunderstorms. The time of maximum frequency of inland thunderstorms is afternoon, the time when the vertical convection is at a maximum. Over the ocean, because of evaporation and the high specific heat of water, the surface temperature rises slowly during the day and falls slowly during the night. The diurnal range of temperature at the ocean surface is much less than that of the air, and hence temperature gradients over the ocean favorable for rapid vertical convection are most frequent during the early morning hours; therefore the maximum frequency of ocean thunderstorms is between midnight and 4 A.M., quite different from the case on land, where the storm occurs in the afternoon.

Time of greatest frequency

Just as thunderstorms occur most frequently during the hottest hours over the land, so they are most frequent in the hottest months over the land. In middle latitudes the maximum frequency occurs in June, and in higher latitudes in July or August. Over the ocean, on the contrary, temperature gradients favorable for the genesis of thunderstorms occur most frequently during the winter. There also appears to be a relation between the rainfall and thunderstorm frequency, a relation which might be expected.

Humphreys advances the view that, "omitting the effects of radiation, there seem to be but three possible ways by which the cooling of a thunderstorm may be obtained: (*a*) by the descent of originally potentially cold air; (*b*) by chilling the air with the cold rain; (*c*) by evaporation. Each of these will be considered separately.

Descent of cold air inadequate

"(*a*) Obviously, no portion of the upper air could maintain its position if potentially, even slightly, colder than that near the surface. If at all potentially colder it would fall until it itself became the surface air, as indeed is the case in all vertical circulation. Hence the great decrease in temperature that comes with a thunderstorm is not the result of the descent of a layer of air originally potentially cold.

"(*b*) Let the under surface of the thunderstorm cloud be 1,500 meters above the earth, and the column of air cooled by the cold rain and its evaporation 2,000 meters high. Let the surface temperature be 303°A., and the temperature gradient before the storm begins adiabatic up to the under-cloud level, and let there be a rainfall of 20 millimeters.

"Now, at the temperature assumed, a column of air 2,000 meters high whose cross-section is 1 square centimeter weighs, roughly, 210 grams, and its heat capacity, therefore, is approximately that of 50 grams of water. At the top of this column the temperature can be, at most, only about 20 degrees lower than at the bottom, and if the rain leaves the top at this temperature but reaches the earth 7 degrees colder than the surface air before the storm (temperatures that seem at least to be of the correct order) it will have been warmed 13 degrees during its fall and the air column cooled on the average about 0.5 degree. But, as a matter of fact, the air usually is cooled by from 5 to 10 degrees. Hence, while the temperature of the air necessarily is reduced to some extent by mere heat conduction to the cold rain, much the greater portion of the cooling clearly must have some other origin. Let us see, then, if evaporation really is adequate to meet these demands.

Chilling due to rainfall inadequate

"(*c*) It is a common thing in semiarid regions to see a heavy shower, even a thunder shower, leave the base of a cloud and yet fail utterly, because of evaporation, to reach the surface of the earth. Hence it appears quite certain that in the average thunderstorm a considerable portion of the rain that leaves the cloud is evaporated before it reaches the ground, and therefore that the temperature decrease of the atmosphere is largely due to this fact. But if so, why, then, one might properly ask, does not an equally great temperature drop accompany all heavy rains?

Effect of evaporation

"The answer is obvious, because, as a rule, the temperature is higher and the relative humidity lower during a thunderstorm than at the time of an ordinary rain. The chief, perhaps the sole, reason for this difference in relative humidity is the difference in the two cases between the movements of

13

the air. In the thunderstorm the descending air, which can be no more than saturated at top, dynamically warms so rapidly and is so continuously renewed that evaporation into it cannot keep pace with its vapor capacity. During other rains, however, where there is no atmospheric descent, and therefore no dynamical heating, approximate saturation must soon obtain; hence but little further evaporation and, of course, but little cooling.

"But no matter how nor to what extent the details may vary, it seems quite certain that the cold rain of a thunderstorm and its evaporation together must establish a local downrush of cold air—an observed important and characteristic phenomenon, really the immediate cause of the vigorous circulation, whose rational explanation has been attempted in the past few paragraphs.

"As the column or sheet of cold air flows down it maintains in great measure its original velocity and, therefore, on reaching the earth rushes forward in the direction of the storm movement, underrunning and buoying up the adjacent warm air. And this condition, largely due, as explained, to condensation and evaporation, once established necessarily is self-perpetuating, so long as the general temperature gradient, humidity, and wind direction are favorable. It must be remembered, however, that thunderstorm convection, rising air just in front and descending air with the rain, does not occur in a closed circuit, for the air that goes up does not return, nor does the air that comes down go up again immediately; there simply is an interchange between the surface air in front of the storm and the upper air in its rear. The travel of the storm, by keeping up with the underrunning cold current, just as effectually maintains the temperature contrast essential to this open-circuit convection as does continuous heating on one side and cooling on the other maintain the temperature contrast essential to a closed-circuit convection.

Underrunning cold current

"The movements of the warm air in front of the rain, the lull, the inflow, and the updraft resemble somewhat those of a horizontal cylinder resting on the earth where the air is quiet and rolling forward with the speed of the storm.

Similarly, the cold air in its descent and forward rush, together with the updraft of warm air, also resembles a horizontal cylinder, but one sliding on the earth and turning in the opposite direction from that of the forward-rolling or all-warm cylinder. In neither case, however, is the analogy complete, for, as above explained, the air that goes up remains aloft while the cold air that comes down is kept by its greater density to the lower levels. The condition of flow persists, as do cataracts and crest clouds, but here, too, as in their case, the material involved is ever renewed.

Condition of flow

"Between the uprising sheet of warm air and the adjacent descending sheet of cold air, horizontal vortices are sure to be formed in which the two currents are more or less mixed. The lower of these vortices can only be *inferred* as a necessary consequence of the opposite directions of flow of the adjacent sheets of warm and cold air, for there is nothing to render them visible. Neither can any vortices that may exist within the cloud be seen. Near the front lower edge of the cumulo-nimbus system, however, and immediately in front of the sheet of rain, or rain and hail, the rising air has so nearly reached its dew point that the somewhat lower temperature, produced by the admixture of the descending cold air, is sufficient to produce in it a light foglike condensation which, of course, renders any detached vortex at this position quite visible.

The squall cloud

"This squall cloud, in which the direction of motion on top is against the storm, may be regarded as a third horizontal thunderstorm cylinder much smaller but more complete than either of the others.

"The above conceptions of the mechanism of a thunderstorm can, perhaps, be made a little clearer with the aid of illustrations. Fig. 67, a schematic picture of a thunderstorm in the making, gives the boundary of a large cumulus cloud from which rain has not yet begun to fall, and the stream lines of atmospheric flow into it. When the cloud is stationary and there is no surface wind the updraft obviously will be more or less symmetrical about a vertical through its center, but when it has an appreciable

Diagram of thunderstorm

velocity, as indicated in the figure, it is equally obvious that most, often nearly all, of the air entering the cloud will do so through its front under-surface. At this stage there will be no concentrated or local down current, only an imperceptible counter settling of the air round about, because, as previously

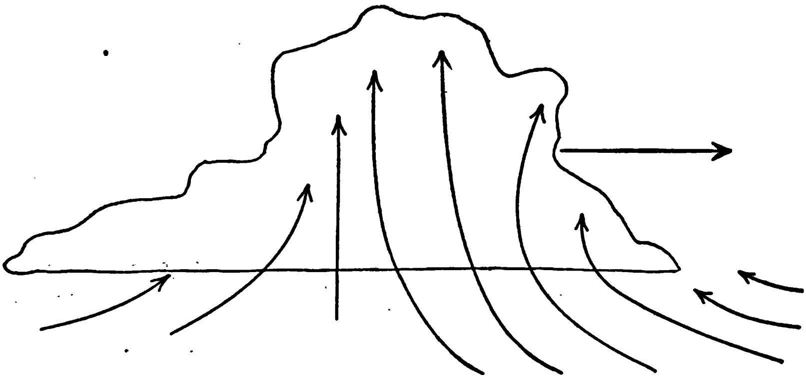


FIG. 67. THUNDERSTORM IN THE MAKING

explained, the air cataract requires local cooling to subpotential temperatures, and this in turn requires local rain.

"Fig. 68 schematically represents a well-developed thunderstorm in progress. The falling rain, often mixed with hail, cools the air through which it falls; and as the temperature gradient was already closely adiabatic it follows that the actual temperatures will be subpotential from the surface of the earth to within the cloud, or throughout and a little beyond the nonsaturated or evaporating levels. As soon, then, as this column or sheet of air is sufficiently cooled it flows down and forward, and all the atmospheric movements peculiar to the thunderstorm are established, substantially as shown.

"Referring to the figure, the warm ascending air is in the region *A;* the cold descending air at *D;* the dust cloud (in dry weather) at *D′*, the squall cloud at *S;* the storm collar at *C;* the thunder heads at *T;* the hail at *H;* the primary rain, due to initial convection, at *R;* and the secondary rain at *R′*. This latter phenomenon, the secondary rain, is a thing of frequent occurrence and often is due, as indicated in the

figure, to the coalescence and quiet settling of drops from an abandoned portion of the cumulus in which and below which winds and convection are no longer active.

"Mammato-cumuli rarely, false cirri frequently, and cap-clouds occasionally, accompany thunderstorms, but as they

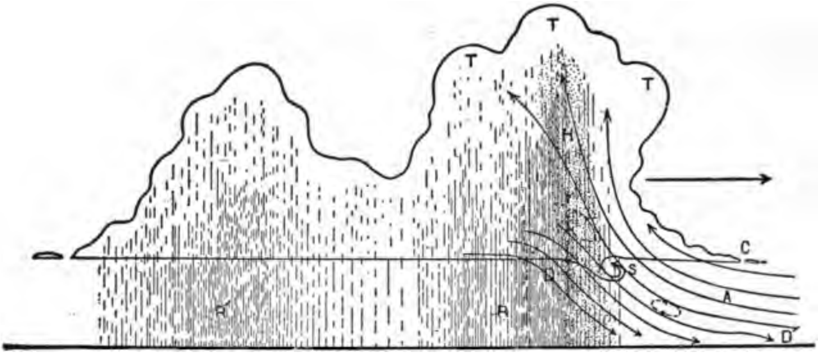


FIG. 68. IDEAL CROSS-SECTION OF A TYPICAL THUNDERSTORM
A, ascending air; D, descending air; C, storm collar; S, roll scud; D′, wind gust; H, hail; T, thunder heads; R, primary rain; R′, secondary rain

are not essential to a thunderstorm they therefore are omitted from the above schematic illustration.

"Before the onset of a thunderstorm there usually, if not always, is a distinct fall in the barometer. At times this fall is extended over several hours, but whether the period be long or short the rate of fall usually is greatest at the near approach of the storm. Just as the storm breaks, however, the pressure rises very rapidly, almost abruptly, usually from 1 to 3 kilobars (1 to 2 millimeters), fluctuates irregularly, and finally, as the storm passes, again becomes rather steady but at a somewhat higher pressure than prevailed before the storm began.

"The cause of these pressure changes is, doubtless, rather complex. The decrease in the absolute humidity and the decrease in temperature both tend to increase the atmospheric pressure, and, presumably, each contributes its share. Both these effects, however, are comparatively permanent, and while they may be mainly responsible for the increase of pressure that persists after the storm has gone by, they probably are not the chief factors in the production of the initial and quickly produced pressure maximum. Here at

least two factors, one obvious, the other inconspicuous, are involved. These are (*a*) the rapid downrush of air, and (*b*) the interference to horizontal flow caused by the vertical circulation.

"The downrush of air clearly produces a vertically directed pressure on the surface of the earth, in the same manner that a horizontal flow produces a horizontally directed pressure against the side of a house. But a pressure equal to a force of 3 kilobars, a pressure increase frequently reached in a thunderstorm, would mean about 3 grams per square centimeter, or 30 kilograms per square meter, and require a wind velocity of roughly 50 kilometers per hour, or 14 meters per second. Now the velocity of the downrush of air in a thunderstorm is not at all accurately known, but while at times probably very considerable, the above value of 14 meters per second seems to be excessive; in fact, its average value may not be even half so great. If in reality it is not, then, since the pressure of a wind varies as the square of its velocity, it follows that less than one fourth of the actual pressure increase can be caused in this way. Hence it would seem that there probably is at least one other pressure factor, and, indeed, such a factor obviously exists in the check to the horizontal flow caused by vertical convection.

"Mere mingling of the two air currents, upper and lower, cannot change the depth of the atmosphere, nor, therefore, the height of the barometer. But in the case of atmospheric convection we have something more than the simple mingling of two air currents, and the linear momentum does not, in general, remain constant. The increased surface velocity following convection, a phenomenon very marked in the case of a thunderstorm, causes an increased frictional drag and therefore a greater or less decrease in the total flow. Suppose this amounts to the equivalent of reducing the velocity of a layer of air only 25 meters thick from V to v, and let $V=5v$. That is, the one three-hundred-and-twentieth part of the atmosphere has its flow reduced to one fifth its former value. This would reduce the total flow by about 1 part in 400, and thereby increase the barometric reading by nearly 3 kilobars.

"It would seem, then, that the friction of the thunderstorm gust on the surface of the earth, through the consequent decrease in the total linear momentum of the atmosphere and, therefore, its total flow, must be an important contributing cause of the rapid and marked increase of the barometric pressure that accompanies the onset of a heavy thunderstorm.

"To sum up: The chief factors contributing to the increase of the barometric pressure during the thunderstorm appear to be, possibly, in the order of their magnitude, (*a*) decrease of horizontal flow, due to surface friction; (*b*) vertical wind pressure, due to descending air; (*c*) lower temperature; (*d*) decrease in absolute humidity.

Change in temperature

"Before the onset of the storm the temperature commonly is high, but it begins rapidly to fall with the first outward gust and soon drops often as much as from five to ten degrees; because, as already explained, this gust is a portion of the descending air cooled by the cold rain and by its evaporation. As the storm passes the temperature generally recovers somewhat, though it seldom regains its original value.

High humidity

"Heavy rain, at least up in the clouds, and therefore much humidity, and a temperature contrast sufficient to produce rapid vertical convection, are essential to the genesis of a thunderstorm. Hence during the early forenoon of a thunderstorm day both the absolute and the relative humidity are likely to be high. Just before the storm, however, when the temperature has greatly increased, though the absolute humidity still is high, the relative humidity is likely to be rather low. On the other hand, during and immediately after the storm, because chiefly of the decrease in temperature, the absolute humidity is comparatively low and the relative humidity high.

Rain gushes

"It has frequently been noted that the rainfall is greatest after heavy claps of thunder, a fact that appears to have given much comfort and great encouragement to those who maintain the efficacy of mere noise to induce precipitation — to jostle cloud particles together into raindrops. The correct explanation, however, of this phenomenon seems obvious: The violent turmoil and

spasmodic movements within a large cumulus or thunderstorm cloud cause similar irregularities in the condensation and resulting number of raindrops at any given level. These in turn, as broken by the air currents, give local excess of electrification and of electric discharge or lightning flash. We have, then, starting toward the earth at the same time and from practically the same level, mass, sound, and light. The light travels with the greatest velocity, about 300,000 kilometers per second, and, therefore, the lightning flash is seen before the thunder is heard. The velocity of sound is only about 330 meters per second. But the rain falls much slower still and therefore reaches the ground after the thunder is heard. In reality it is the excessive condensation or rain formation up in the cumulus cloud that causes the vivid lightning and the heavy thunder. According only to the order in which their several velocities cause them to reach the surface of the earth it might appear, and has often been so interpreted, that the lightning, first perceived, was the cause of the thunder, which, indeed, it is, and that the heavy thunder, next in order, was the cause of the excessive rain.

Velocity of thunder and raindrops

"The velocity of the thunderstorm is simply the velocity of the atmosphere in which the bulk of the cumulus cloud happens to be located. Hence, as the wind at this level is faster by night than by day and faster over the ocean than over land, it follows that exactly the same relations hold for the thunderstorm, that it travels faster over water than over land and faster by night than by day. The actual velocity of the thunderstorm, of course, varies greatly, but its average velocity in Europe is 30 to 50 kilometers per hour; in the United States, 50 to 65 kilometers per hour.

Rate of movement of storms

"Hail, consisting of lumps of roughly concentric layers of compact snow and solid ice, is a conspicuous and well-known phenomenon that occurs during the early portion of most severe thunderstorms. But in what portion of the cloud it is formed and by what process the layers of ice and snow are built up are facts that, far from being

Hail

obvious, become clear only when the mechanism of the storm itself is understood.

"As before, let the surface temperature be 303° A. and the absolute humidity 40 per cent, or the dew point 288° A. Under these conditions saturation will obtain, and, therefore, cloud formation will begin when the surface air has risen to an elevation of 1.5 kilometers. Immediately above this level the latent heat of condensation reduces the rate of temperature decrease with elevation to about half its former value, nor does this rate rapidly increase with further gain of height. Hence, usually, for the above assumptions correspond in general to average thunderstorm conditions, it is only beyond the 4-kilometer level that freezing temperatures are reached. It is therefore only in the upper portions of cumulus clouds, the portions that clearly must consist of snow particles and undercooled fog or cloud droplets, that hail can either originate or greatly grow.

"But what, then, is the process by which the nucleus of the hailstone is formed, and its layer upon layer of snow and ice built up? Obviously such drops of rain as the strong updraft within the cloud may blow into the region of freezing temperatures will quickly congeal and also gather coatings of snow and frost. After a time each incipient hailstone will get into a weaker updraft, for this is always irregular and puffy, or else will tumble to the edge of the ascending column. In either case it will then fall back into the region of liquid drops, where it will gather a coating of water, a portion of which will at once be frozen by the low temperature of the kernel. But again it meets an upward gust, or falls back where the ascending draft is stronger, and again the cyclic journey from realm of rain to region of snow is begun; and each time—there may be several—the journey is completed a new layer of ice and a fresh layer of snow are added. In general, the size of the hailstones will be roughly proportional to the strength of the convection current, but since their weights vary approximately (they are not homogeneous) as the cube of their diameters while the supporting force of the upward air current varies, also approximately, as only the square of their diameters, it follows that a limiting size is

quickly reached. It is also evident, from the fact that a strong convection current is essential to the formation of hail, that it can occur only where this convection exists; that is, in the front portion of a heavy to violent thunderstorm.

"Some meteorologists hold that the roll scud between the ascending warm and descending cold air is the seat of hail formation, but this is a mistaken assumption. Centrifugal force would throw a solid object, like a hailstone, out of this roll probably before a single turn had been completed. Besides, and this objection is, perhaps, more obviously fatal than the one just given, the temperature of the roll scud, because of its position, the lowest of the whole storm cloud, clearly must be many degrees above the freezing point. Indeed, temperatures low enough for the formation of hail cannot obtain at levels much less than three times that of the scud. Therefore, it is clearly in the higher levels of the cumulus and not in the low scud that hail must have its genesis and make its growth."

Hail not due to roll scud

Walter

FIG. 69. THE GROWTH OF AN ELECTRIC SPARK DISCHARGE

49. Lightning. Dr. Walter of Hamburg has obtained, by means of a rotating camera, proof that lightning flashes are not, as heretofore generally thought to be, oscillatory discharges, but mainly unidirectional. A lightning flash begins with a preliminary branching spark followed within a brief interval, say .01 second, by another rupture or discharge somewhat longer, until finally a path is built up for a main-line discharge, which again is intermittent. According to this view, a discharge is something like a tearing or ripping, but, of course, done in a very brief interval of time. Many photographs, however, serve to indicate this progressive character of

Lightning flashes not oscillatory discharges

the main line of breakdown of the dielectric, or air. (Figs. 69, 70, and 71.)

Walter

FIG. 70. STREAK LIGHTNING, STATIONARY CAMERA
Companion to FIG. 71

Dr. Wilhelm Schmidt has pointed out that, because of the great accumulation of energy in a very small space, there must be a strong repulsion of electrified air particles of like sign and a sudden increase of pressure in the path of the discharge. This may amount to perhaps 100 atmospheres. A wave, therefore, of intense energy spreads out in all directions—an explosion wave. There is a strong push or condensation, and rebound or rarefaction succeeded by waves of less intensity. For studying the wave motion during thunderstorms, Schmidt used two forms of apparatus, one for analysis of regular sound waves and the other for the longer pressure waves. The records showed that regular trains of waves of uniform length practically never occurred and hence that thunder had no proper "tone," but was merely a "noise" similar in character to the rattling of window panes. Irregularity was greatest during the heaviest thunder, while near the end of the peal a certain regularity was noticed. A statistical analysis of the frequency with which waves of various lengths occurred showed a

Electrical discharge an explosion wave

Walter

FIG. 71. STREAK LIGHTNING (SEQUENT DISCHARGE), ROTATING CAMERA
Companion to Fig. 70

preponderance of those lasting .025 second or more, and again of those lasting between .0083 and .013 second, that is, vibrations of such length that if they had occurred in uniform trains they would have produced the tone E or lower, or again tones between D and A. Shorter waves most common in music were rarer.

The fluctuations in air density occurring with these waves were larger than in ordinary sound waves. Although the lightning was never close at hand, the interval between flash and beginning of thunder being about 5 seconds, the pressure fluctuations were, as a rule, more than 1.3 kilobars (1 millimeter). "Hence," says Schmidt, "the greater part of the total energy of thunder is represented by long, inaudible waves, and, strange as the statement may sound, we may say that one really hears only the smallest part of a clap of thunder. Most of the phenomenon either escapes our senses altogether or is perceptible only through the vibration of objects around us, as the rattling of window panes. In the immediate vicinity of the discharge the pressure fluctuations must be very violent, and much of the purely mechanical injury wrought by lightning may be due to them."

Only a small part of a clap of thunder audible

The number of violent waves is small, occurring generally in irregular series of three or four. In the heaviest thunder claps there is usually but one violent wave,— at the beginning. Such a wave, called a "shock wave," travels in all directions from the path of the electrical discharge. The prolongation of the sound is due to the fact that the discharge is perhaps intermittent and may set up several initial waves; also because of reflection, not so much from clouds and sheets of falling rain as from the "interfaces" between atmospheric strata of different temperatures—and especially by the action of the wind. The original sharp report is transformed into a "roll." Irregular short waves, which give the rattling noise of near-by thunder, are gradually lost in the more regular waves so that in distant thunder the sound may have a definite pitch. The energy of thunder as shown by these records amounted in a maximum case to 22,000 kilogrammeters, and was therefore

"Shock wave"

very great compared with that of ordinary sounds. In this case the thunder lasted 13 seconds, and it would require more than 200,000,000 buglers, blowing for the same length of time, to produce an equivalent amount of energy. Nevertheless, this quantity is insignificant compared with that of a flash of lightning, for which we may assume, and not in extreme cases, something like 10,000,000,000 kilogrammeters. In fact, only a small part of the energy of lightning is transformed into pressure waves and sound, most of it assuming other forms, such as those of heat or light.

Tremendous energy of thunder

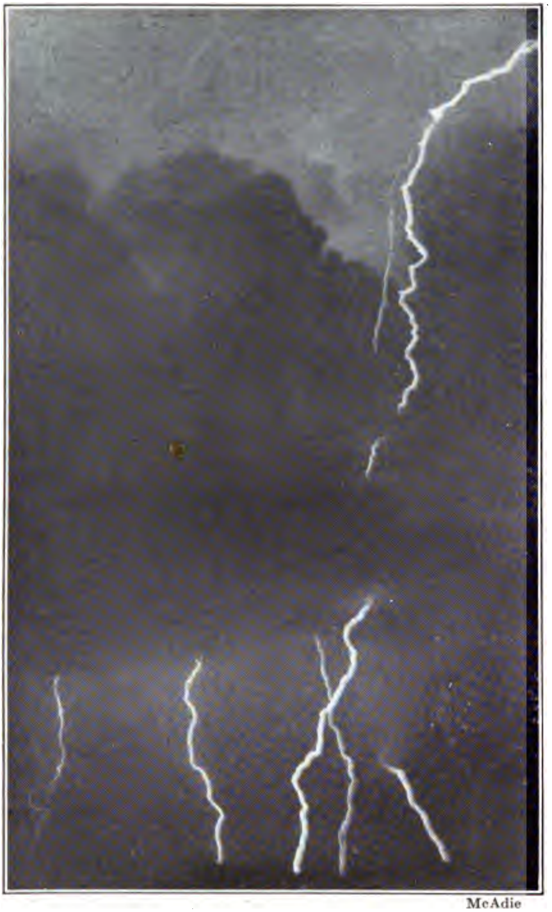
With regard to the explanation of the "rolling" of thunder as due to the fact that the sound reaches the observer first from the nearest part of the path and later from the more distant parts, and that the duration of thunder depends upon these intervals,— Schmidt points out that in the case of a uniform impulse along a path free from sharp angles there would be only a single wave, spreading in all directions, and that the observer would therefore perceive only a single brief sound, and that the time of occurrence would depend upon the nearest part of the lightning path. Bends in the lightning path will account only for a limited number of "claps," and not for the "roll" of thunder.

The "rolling" of thunder

It has been suggested by Miessner that there may be electrolytic and thermal decomposition of the water vapor and subsequent explosion by the lightning. Further experimentation is necessary.

L. A. De Blois has made special investigations on the character of the lightning discharge. Of fourteen electrical storms in the summer of 1913, six were of a character suitable for his observations, which were made primarily to determine suitable protection for buildings containing explosives.[1] He found that after some experience with the sound produced in wireless receivers by waves propagated by lightning discharges, he could predict the occurrence of the storm by eight or ten hours. His instruments were a wireless receiving outfit with

[1] De Blois' results are published in a paper presented before the 294th meeting of the American Institute of Electrical Engineers, April 24–25, 1914. A criticism by C. F. Marvin of the deductions drawn from these experiments can be found in the *Monthly Weather Review* for Aug., 1914.



McAdie

FIG. 72. LIGHTNING DISCHARGE THROUGH CLOUDS

various detectors, an indicating ceraunoscope, a static voltmeter, and an oscillograph, which, when in circuit, discharged at a frequency of about 400,000 oscillations per second. The natural period of the oscillograph itself was only about 5,000. De Blois gives three characteristic lightning oscillograms. The first was a typical example of a single steep-front discharge from clouds to earth. The maximum value of the induced current was 0.5 ampere, and the peak proper endured 0.0008 second. The second was a typical steep-front wave with five main and three supplementary peaks. The third was a wave of totally different character. The direction of flow is from the earth upward, but with a gradual rise, with indications of oscillations to a maximum of 0.18 ampere, which is reached in 0.0046 second. De Blois sums up his photographic records of 50 discharges as follows:

Total number of strokes positive (from clouds to earth), 43;
Total number of strokes positive (from earth to clouds), 7;
Single-peak discharges, 38, average duration 0.00065 second;
Multiple-peak discharges, 12.

De Blois states that in no record was there any indication of regular periodic high-frequency oscillations in the induced current. The peaks may represent simply the result of the "progressive breakdown" of the atmosphere, defined by Steinmetz; that is, a discharge from point to point. Steinmetz has likened the discharge of a cloud to a landslide which sets off a series of landslides. He asks us to imagine a relief map of wet sand, the hills representing the dense portion of the cloud or the places of high potential, and the valleys the places of light, or low, potential. Where the declivity is very steep a slide occurs which causes another slide and so on until the hills are leveled and the valleys filled; or, in other words, until the electric potential is equalized. Whatever may be the true explanation, it would appear that the most dangerous discharges are those with almost instant rise to maximum value.

Humphreys states that "curious luminous balls or masses, of which C. De Jans[1] probably has given the fullest account, have time and again been reported among the phenomena observed during a thunderstorm. Most of them appear to last only a second or two and to have been seen at close range, some even passing through a house, but they have also appeared to fall, as would a stone or like a meteor, from the storm cloud, and along the approximate path of both previous

McAdie

FIG. 73. LIGHTNING FLASH
Note the dark flashes.

[1] *Ciel et terre*, Bruxelles, Vol. XXXI (1910), p. 499.

Photograph by A. Steadworthy, Dom. Astr. Obs. Ottawa, in Jour. of the Royal Astr. Soc. of Canada, 1912

FIG. 74. PHOTOGRAPH OF LIGHTNING TAKEN IN DAYLIGHT, JULY 10, 1912

and subsequent lightning flashes. Others appear to start from a cloud and then quickly return, and so on through an endless variety of places and conditions.

"Doubtless many reported cases of ball lightning are entirely spurious, being either fixed or wandering brush discharges or else nothing other than optical illusions, due in most cases probably to persistence of vision. But here, too, as in the case of rocket lightning, the amount and excellence of observational evidence forbid the assumption that all such phenomena are merely subjective. Possibly in some instances, especially those in which it is seen to fall from the clouds, ball lightning may be only extreme cases of rocket lightning, cases in which the discharge for a time just sustains itself. A closely similar idea has been developed in

detail by Toepler. The ball may either disappear wholly and noiselessly, as often reported, or it could perhaps suddenly gain in strength and instantly disappear, as sometimes observed, with a sharp, abrupt clap of thunder.

"To say that all genuine cases of ball lightning, those that are not mere optical illusions, are stalled thunderbolts certainly may sound very strange. But that, indeed, is *just* what they are according to the above speculation—a speculation that recognizes no difference in *kind* between streak, rocket, and ball lightning, only differences in the *amounts* of ionization, quantities of available electricity, and steepness of potential gradients.

Ball lightning similar to rocket type

"When a distant thundercloud is observed at night one is quite certain to see in it beautiful illuminations, looking like great sheets of flame, that often flicker and glow in exactly the same manner as does streak lightning for well-nigh a whole second. In the daytime and in full sunlight the phenomenon when seen at all appears like a sudden sheen that travels and spreads here and there over the surface of the cloud. Certainly in most cases—so far as definitely known in all cases—this is only reflection from the body of the cloud of streak lightning in other and invisible portions. Conceivably a brush or coronal discharge may take place from the upper surface of a thunderstorm cloud, but one would expect this to be either a faint continuous glow, or else a momentary flash coincident with a discharge from the lower portion of the cloud to earth or to some other cloud. But, as already stated, only reflection is definitely known to be the cause of sheet lightning. Coronal effects seem occasionally possible, but that they are ever the cause of the phenomenon in question has never clearly been established and appears very doubtful. It has often been asserted, too, that there is a radical difference between the spectra of streak and sheet lightning, but even this does not appear ever to have been photographically proved."

Sheet lightning

Glow discharge

50. Other forms of discharge. These are *beaded* lightning, or apparently discontinuous streaks; *return* strokes; *dark* flashes (Fig. 73), which are believed to be photographic effects; *heat* or distant discharges; *St. Elmo's fire;* and *rocket* lightning.

"The spectrum of a lightning flash and that of an ordinary electric spark in air are practically identical. This is well shown by Fig. 75, copied from an article on the spectrum of lightning by Fox,[1] in which the upper or wavy portion is due to the lightning and the lower or straight portion to a laboratory spark in air."

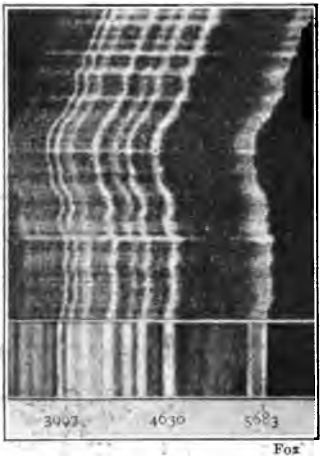


FIG. 75. SPECTRUM OF LIGHTNING

A. Steadworthy gives the following description[2] of how he secured a spectrum of lightning:

"A favorable opportunity offered on the night of July 10, 1912, when a storm began to develop at 6 P.M. at Ottawa in the west. Low-lying cumulus clouds slowly spread toward the north and south where distant lightning first manifested itself. The clouds, becoming denser and denser, approached Ottawa and slowly rose above the horizon. By 7 P.M. the first rumblings of thunder began, and by 9:30 P.M. the pyrotechnic display in the west became so vivid that I brought out the two cameras which I had in readiness at my house overlooking McDonald Park, giving me an outlook from the west through the north to the east. One camera was fitted with stereoscopic lenses, the other was an 8-inch by 10-inch camera with 2-inch lens, 16-inch focus. In front of it was fitted in a box one of our 60-degree dense flint-glass objective prisms. The box could be rotated on the collar of the lens, thereby enabling spectra to be obtained over a range of 180 degrees without moving the camera. It must be observed here that the size of the small box and the prism not physically covering the lens made it possible for a flash coming from a certain direction to get through the lens and on to the plate without passing through the prism. And this is exactly what happened, as seen in Fig. 76, which shows the spectrum secured with the superimposed other flash.

[1] *Astrophysical Journal*, Chicago, Vol. XVIII (1903), p. 294

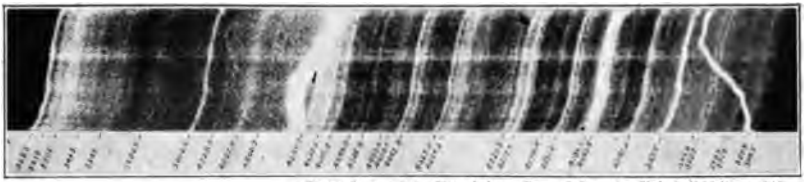
[2] *Jour. of the Royal Astr. Soc.*, Canada, Nov.–Dec., 1912.

Courtesy of Dr. C. A. Chant
From Jour. of the Royal Astr. Soc. of Canada, Vol. 8 (1914), p. 346

FIG. 76. SPECTRUM OF LIGHTNING

This photograph was taken at Ottawa by A. Steadworthy, July 10, 1912. The straight streaks at the bottom are spectra of street lamps.

"I made four exposures on Cramer isochromatic plates, for, as I thought, four spectra. One of the plates turned out a



From Jour. of the Royal Astr. Soc. of Canada, Vol. 8 (1914), p. 345

FIG. 77. SPECTRUM OF LIGHTNING

This is a portion of Fig. 76, with wave-lengths added. It is magnified to five times the original.

blank, one showed a very weak spectrum, one showed the fine spectrum on the above plate, while the last showed remarkable dotted curved lines, one at the top of the plate and several parallel ones below. I utterly fail to divine their origin. The exposure and the position of the camera were identical in the two cases. None of the other three plates shows this phenomenon."

LINES IN SPECTRUM OF LIGHTNING[1]

Wave length	Character	Fox's wave length†	
5683.0*	Strong, well-defined line	5683	r. edge
5618.3	Broad	5600	max.
5556.3	Broad and strong		
5423.2	Violet edge of gradually fading band		
5345.5	Faint, broad band	5306	v. edge
5180.9	Faint, broad line	5175	
5003.7*	Good, strong band	5003.7	
4928.5	Faint line		
4852.4	Broad, fairly strong band	4842	
4800.7	Faint line	4786	
4688.5	Fairly good line	4660	r. edge
	Faint indistinct lines between 4688.5 and 4648.9		
4648.9	Faint line		
4635.6	Strong, clear line	4630.7	
	More faint lines		
4605.3	Fairly strong broad lines	4603	v. edge

[1] By A. Steadworthy and J. B. Cannon, Dominion Observatory, Ottawa, Canada, Sept., 1914.

LINES IN SPECTRUM OF LIGHTNING—Continued

Wave length	Character	Fox's wave length†
4542.8	Faint line	4535
4522.8	Fairly strong line	
4483.8 / 4468.6	A pair of rather weak lines, v. one the stronger	
4442.5	Strong, clear line	4439
4427.8	Line fainter and broader	
4413.4	Very faint line	
4367.9	Strong line	4359
4354.3	Line less strong	
4340.9	Fainter, broad line	
4308.1	Faint, broad band	
4270.2	Very faint	
4264.1	Very faint	
4251.9	Faint line	
4239.8	Strong line	4238
4217.7	Strong, clear line	
4216.1	Faint line	
4190.1	Very faint line	4183
4170.4	Fairly strong line	
4151.2	Good, strong line	4154
4143.1	Good line, not so strong as 4151	
4127.0	Very faint line	
4106.1 / 4095.8	Very strong line, forms a close strong pair with 4095.8	4105
4070.6	Rather faint	4077
4041.3	Good strong line	4041.5
4036.5	Faint line	
4024.7	Faint line	
4015.3	Faint line	
3997.0*	Good, strong line	3997
3959.2 / 3952.7	Faint pair	3943
3925.2 / 3919.0	Faint pair	3915
3898.7	Faint line	3890
3888.7	Very faint	3838

*Lines marked thus were used as standards in the compilation of wave lengths.
†*Astrophysical Journal*, Chicago, Vol. XVIII (1903), p. 294.

51. Destructive effects of lightning. One of the most striking effects of lightning is the shattering of objects struck. The discharge apparently prefers the surface and the effects are largely superficial. Thus the clothes are sometimes torn from the body, splinters of flagpoles are thrown great distances, and it has happened that where three men have been standing under a tree, the one in the center of the group was only injured, while the man on his right and the man on his left were killed. Such an occurrence was reported at Lowell, Mass., on July 8, 1916. Humphreys explains the explosive effect as follows:

Explosive effects

"The excessive and abrupt heating caused by the lightning current explosively and greatly expands the column of air through which it passes. It even explosively vaporizes such volatile objects as it may hit that have not sufficient conductivity to carry it off. Hence, chimneys are shattered, shingles torn off, trees stripped of their bark or utterly slivered and demolished, kite and other wire fused or volatilized, holes melted through steeple bells and other large pieces of metal, and a thousand other seeming freaks and vagaries wrought.

"Many of the effects of lightning appear at first difficult to explain, but, except the physiological and, probably, some of the chemical, all depend upon the sudden and intense heating along its path.

Chemical effects

"As is well known, oxides of nitrogen and even what might be termed the oxide of oxygen, or ozone, are produced along the path of an electric spark in the laboratory. Therefore, one might expect an abundant formation, during a thunderstorm, of these same compounds. And this is exactly what does occur, as observation abundantly shows. It seems probable, too, that some ammonia must also be formed in this way, the hydrogen being supplied by the decomposition of raindrops and water vapor.

"In the presence of water or water vapor these several compounds undergo important changes or combinations. The nitrogen peroxide (most stable of the oxides of nitrogen) combines with water to produce both nitric and nitrous acids; the ozone with water gives hydrogen peroxide and sets oxygen free; and the ammonia in the main merely dissolves, but

probably also to some extent forms caustic ammonia and hydrogen.

"The ammonia and also both the acids through the production of soluble salts are valuable fertilizers. Hence, wherever the thunderstorm is frequent and severe, especially, therefore, within the tropics, the chemical actions of the lightning may materially add, as has recently been shown,[1] to the fertility of the soil and promote the growth of crops.

Normal atmospheric electricity

"The only reason for mentioning normal atmospheric electricity in connection with thunderstorms is to emphasize the fact that, contrary to what many suppose to be the case, there is but little relation, in the sense of cause and effect, between these two phenomena. Thus while the difference in electrical potential between the surface of the earth and a point at constant elevation is roughly the same at all parts of the world, the number and intensity of thunderstorms vary greatly from place to place.

Potential gradient and thunderstorm

Further, while the potential gradient at any given place is greatest in winter, the number of thunderstorms is most frequent in summer, and while the gradient, in the lower layer of the atmosphere, at many places, usually is greatest from 8 to 10 o'clock, both morning and evening, and least at 2 to 3 o'clock P.M. and 3 to 4 o'clock A.M., no closely analogous relations hold for the thunderstorm."

However, it should be noted that there are marked fluctuations in the potential values during thunderstorms and snow storms. Characteristic curves are those shown in Figs. 78 and 79 obtained by the author in experiments made at the top of the Washington Monument and at Blue Hill Observatory.

"Probably the most interesting conclusion in regard to normal atmospheric electricity so far drawn from observation and experiment is this: that the earth everywhere, land and water and from pole to pole, is a negatively charged sphere of practically constant surface density, surrounded by an atmosphere of such conductivity that it is constantly carrying away a current that amounts on the whole to about 1,000 amperes.

[1] Capus in *Ann. de Géographie*, Paris, Vol. XXIII (1914), p. 109.

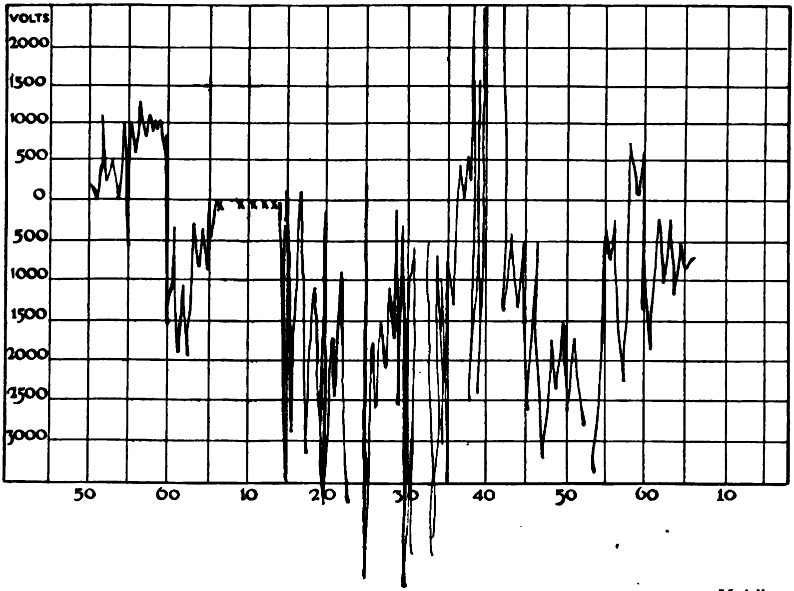


FIG. 78. VOLTAGE OF AIR DURING A THUNDERSTORM AT THE TOP OF THE WASHINGTON MONUMENT

"Where the supply of negative electricity comes from that keeps the surface of the earth on the whole negatively charged in spite of this steady great loss, or how the outgoing current is compensated, no one knows. Rain does not help matters, for, as we have seen, rain is prevailingly positive, and what is needed, to compensate the loss, is negative electricity and a great deal of it. Neither, so far as known, is negative electricity supplied by means of the lightning, for, in the great majority of cases, this, too, is positive that descends from cloud to earth. And so the puzzle remains. As Simpson[1] puts it: "'A flow of negative electricity takes place from the surface of the whole globe into the atmosphere above it, and this necessitates a return current of more than 1,000 amperes; yet not the slightest indication of any such current has so far been found, and no satisfactory explanation for its absence has been given."

Detailed descriptions of methods of obtaining curves of potential may be found in the *Monthly Weather Review*, July,

[1] *Nature*, London, Vol. XC (1912), p. 411.

1891, p. 171; also in *Third Memoir National Academy of Sciences*, Vol. V, by T. C. Mendenhall; also in *Luftelectrizität*, by Karl Kaehler, Berlin, 1913.

In the experiments on atmospheric electricity made by Hewlett, Kidson, and Johnston of the Department of Terrestrial Magnetism of the Carnegie Institution, the most striking result appears to be that, while the conductivity over the ocean is, on the average, at least as great as that over land, the radioactive content is much smaller. The mean value of the potential gradient near the surface of the water is approximately 120 volts per meter.

There appears to be a relation between conductivity and temperature and pressure, but no relation to absolute humidity. An elaborate discussion of the observations made on the second cruise of the *Carnegie* is given by Hewlett in *Terrestrial Magnetism*, September, 1914, p. 127. Specific conductivity, potential gradient, and radioactivity observations were continued for more than two years in the Pacific, Atlantic, and Indian oceans. The potential gradient over the ocean is of the same order of magnitude as over the land.

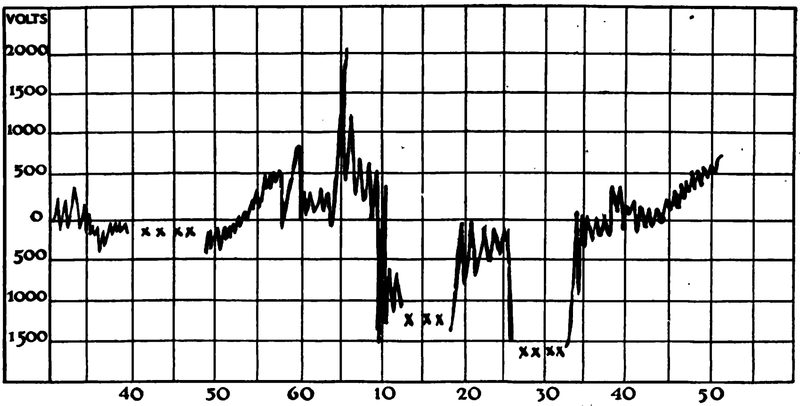


FIG. 79. VOLTAGE OF AIR DURING A SNOWSTORM McAdie

Swann in the same journal discusses new methods and instruments for measuring the specific numbers and velocities of the atmospheric ions, the radioactive content of the atmosphere, and the potential gradient.

52. Protection from lightning. Few questions have been more debated than the certainty of protection afforded by lightning rods and protective devices. We may sum up the controversy in the words of Kelvin, "that there is good reason to feel that there is a very comfortable degree of security, if not of absolute safety, given to us by lightning conductors made according to present and orthodox rules."

Lightning rods

Detailed instructions for proper installation of rods and making good grounds are given in numerous books, among which may be mentioned *The Report of the Lightning Rod Conference*, Anderson on *Lightning Rods*, Lodge on *Lightning Conductors;* and McAdie and Henry in various Weather Bureau publications; also *Bulletin No. 56*, Bureau of Standards.

An interesting statistical study of the ratio of buildings with lightning rods struck to those not struck is that made by J. W. Smith, who ascertained from insurance companies that during 1912 and 1913 of 1,845 buildings struck only 67 were supplied with rods or protective devices. Approximately 31 per cent of the buildings insured were rodded, hence the number of buildings struck should have been 572 instead of 67, or on these figures the efficiency of the rods would be about 90 per cent. Five companies with 18,000 buildings insured, of which 50 per cent were rodded, reported that no building with rods had been burned or even materially damaged. Again, it is noteworthy that when a rodded building is struck the damage is much less than in the case of an unrodded building.

Efficiency of lightning rods

Briefly, it may be said that all barns and exposed buildings should have well-grounded lightning rods; but as for ordinary dwelling houses in city blocks there is little likelihood of their being struck, hence rods may be dispensed with. The nature of a locality determines somewhat the need; thus if the building is near a water course, the risk is greater. Places separated by a few miles have different frequencies of stroke. Lightning may strike several times in the same place; there is no known reason why it should not. It is unwise to stand under or near trees during thunderstorms. Some trees, such as the locust

Barns, spires, and exposed houses should have rods

and the pine, appear to be more frequently struck than others, owing possibly to the character of the root system. However, all trees may be struck.

PERCENTAGE OF DIFFERENT SPECIES STRUCK BY LIGHTNING IN EUROPE

Kind of tree	Per cent	Kind of tree	Per cent
Poplar	30.5	Horse chestnut	0.3
Oak	20.7	Plum	.3
Conifers	15.5	Alder	.2
Elm	4.6	Larch	.2
Pear	3.9	Service	.2
Willow	3.5	Catalpa	.1
Beech	2.4	Elder	.1
Ash	1.8	Lime	.1
Linden	1.8	Maple	.1
Walnut	1.3	Vine	.1
Cherry	1.2	Others	.4
Apple	1.0	Unknown	8.0
Chestnut	.8		
Birch	.5		100.0
Acacia	.4		

Plummer has discussed the liability of trees to lightning stroke[1] and shows that:

Why trees are struck

1. Trees are the objects most often struck by lightning, because (*a*) they are the most numerous of all objects; (*b*) as a part of the earth's surface they extend upward and shorten the distance to a cloud; (*c*) their spreading branches in the air and spreading roots in the ground present the ideal form for conducting an electrical discharge to the earth.
2. Any kind of tree is likely to be struck by lightning.
3. The greatest number of any species struck in any locality is the dominant species of that locality.
4. The likelihood of a tree being struck by lightning is increased (*a*) if it is taller than surrounding trees; (*b*) if it is isolated; (*c*) if it is upon high ground; (*d*) if it is well (deeply) rooted; (*e*) if it is the best conductor at the moment of the flash; that is, if temporary conditions, such as being wet by rain, transform it for the time from a poor conductor to a good one.
5. Lightning may bring about a forest fire by igniting the tree itself, or the humus at its base. Most forest fires caused by lightning probably start in the humus.

[1] *Bulletin No. 111*, Forest Service.

Data have been gathered both in Europe and in the United States on the frequency of lightning stroke upon trees growing in different soils. An average of all results shows the relative percentages to be:

	Per cent
Loam	42
Sand	22
Clay	19
Others, including rock, marl, and calcareous formations	17
	100

This comparison does not take into account the relative areas of the different soils covered in the investigations. A fairer estimate, made in Belgium, and reduced to actual frequency per unit area, gives the following:

	Per cent
Loam	23
Sand	18
Clay	17
Others	42
	100

The frequency of stroke upon the poplar growing in these soils was:

	Per cent
On loamy soil	28
On sandy soil	24
On clay soil	6
On other soils	42
	100

Resuscitation from lightning stroke

The effects of lightning are so remarkable that it is always advisable to attempt to restore consciousness to a person who has been struck, even if the case appears hopeless. There are many cases on record proving the wisdom of such action. There is good reason for believing that in the majority of cases lightning causes suspended animation rather than somatic death. Every effort, therefore, should be made to stimulate respiration and restore circulation. Do not cease in the effort for at least an hour; and be sure to secure the services of a physician as promptly as possible.

CHAPTER XV

Precipitation

RAIN

53. The process that makes the raindrop. One gram of pure water at 277° A., the temperature of maximum density, occupies one cubic centimeter of space. If we change this volume of water into vapor at 373° we find that it expands and now occupies a volume of 1,698 cubic centimeters. When expansion took place, work was done. A reverse process would be the following: If 1,698 cubic centimeters of vapor at the given temperature were compressed into a drop of water one cubic centimeter in volume, or about four times the size of a very large raindrop, a large quantity of heat would be available and there would still remain in the drop considerable internal energy. Of course rain is seldom the result of condensation at so high a temperature.

Compression

Raindrops are not made of pure water. Indeed, it is doubtful whether, under natural conditions, we could expect to find a pure drop, and therefore some allowance must be made in applying the results of laboratory experiments in studies of condensation and evaporation. Furthermore, it is seldom that we find complete saturation. Under natural conditions, even in dense and wet clouds, the true degree of saturation does not exceed 95 per cent. And this explains why it is so hard to dry out a mixed atmosphere and leave not a trace of vapor.

Raindrops not pure water

A molecule of water is approximately one ten-millionth of a millimeter in diameter. The size of a fine drop of rain is from one one-hundredth of a millimeter to half a millimeter in diameter; of a common drop, from 2 to 4 millimeters; of a large drop, from 5 to 7 millimeters.

Size of raindrops

The weight of the largest raindrop does not exceed 0.2 gram, and its diameter is about 7 millimeters. Drops of larger size, according to the experiments of Weisner, Lenard,

and others, when allowed to fall from a height of 22 centimeters separate into small drops. Wilson A. Bentley, by allowing raindrops to fall into a layer 25 millimeters deep of fine uncompacted flour, with a smooth surface, contained in a shallow tin receptacle about 100 millimeters in diameter, exposed to the rain for about four seconds, obtained dough pellets which were found by experiment to correspond very closely in size with the raindrops which made them. These pellets were photographed; and the type of storm, temperature, and approximate height of clouds noted and recorded. Three hundred and forty-four sets of raindrop impressions were secured in this way from 70 different storms. A remarkable discovery made in these investigations was the astonishing differences in the dimensions of the individual drops both in the same and in different rainfalls. While the larger ones possessed diameters of 6 or 8 millimeters, the smaller ones were only 0.8 or 0.6 millimeter, and there were frequently microscopic drops too small to make an impression. The following table shows the relative frequencies. Of a total of 867 drops, there were:

149 very small drops (less than 0.8 mm.), or 17 per cent of total;
288 small drops from 0.8 to 1.5 mm., or 34 per cent of total;
254 medium drops from 1.6 to 3.5 mm., or 29 per cent of total;
141 large drops from 3.6 mm. to 5.1 mm., or 16 per cent of total;
35 very large drops above 5.2 mm., or 4 per cent of total.

Variation with storm locus

Bentley, in the same manner, studied the distribution in 51 storm areas and found that, in general, the very small drops increase in number from the east edge toward the west, or receding edge, of a storm; that for other sizes there was a progressive increase toward the center of a storm, but a decrease toward the west. This law, however, does not hold for thunderstorms. It seems to be a fact that, in general, showers renew themselves or acquire new vigor within the western, or receding, segment. Bentley advances the view that, ordinarily, the size of each individual raindrop depends largely upon and increases with the square of the mass of upper and intermediate clouds that a drop passes through on its journey earthward. Practically all of the rainfall from low and

intermediate clouds (low-lying cumulus, nimbus, cirro-stratus, and cirro-cumulus) consists of small and medium drops. The larger raindrops are shed in considerable numbers only from thick, vertically expanding, complex clouds. It was often noted that whenever electrical discharges were unusually frequent and powerful the rainfall was unusually thick and heavy and consisted of raindrops of all sizes, the larger sizes predominating. The following table compares the sizes of drops that fall during lightning and when there is no lightning:

Heavy rainfall with electrical discharge

	Near lightning	Distant lightning	No lightning
Very small drops..	15	31	38
Small drops.......	61	67	70
Medium drops....	64	63	57
Large drops.......	58	23	22
Very large drops ..	23	3	1

Bentley thinks that the major portion of the rainfall of thunder showers is of snow origin; also that a very large portion of the rainfall in all climates is due to the melting of snowflakes or granular snow.

54. **The rain gauge.** The invention of the rain gauge is attributed to an Italian, Benedetto Castelli, who in June, 1639, informed Galileo that he had measured the rainfall with a vase one *spanne* in depth and half a *spanne* in diameter. But Dr. Y. Wada, director of the Korean Meteorological Observatory, has shown that rain gauges were in use in Korea[1] as early as 1442.[2] The dimensions of the ancient gauge were: depth, 30 centimeters; diameter, 14 centimeters. A gauge of the period of 1770 was found by Wada on an inspection trip, actually in use at the observatory of Chemulpo. On the granite pillar of the gauge represented in the illustration (Fig. 80) found at Taiku are engraved three large Chinese characters meaning "Instrument to measure rain," and seven smaller characters meaning "Constructed in the fifth month of the cycle of the year," a date in the Chinese calendar corresponding to 1770.

Early Korean rain gauges

[1] The modern name for Korea, now a department of Japan, is Chosen.

[2] *Jour. of the Met. Soc. of Japan*, Mar., 1910; also *Quart. Jour. of the Royal Met. Soc.*, Jan., 1911.

From Quart. Jour. of the Royal Met. Soc., 1911

Fig. 80. The Oldest Rain Gauge
This rain gauge was erected at Taiku in 1770.

This is evidence that rain observations began in Korea two centuries before they were made in Europe, and, what is more remarkable, there appears to have been an udometric survey, or network of rain-reporting stations. The reports were probably used in connection with the rice crop. A large amount of water is required in the cultivation of this crop, especially in the first period of its transplanting; and should the rainy season not commence until the middle of July, and the rainfall be deficient, the crop may be injured. It is said that during droughts from the earliest times the rulers of Korea caused prayers for rain to be offered to the saints of the mountains and rivers. This is a striking illustration of the dependence of a community's welfare upon rainfall. Dr. Wada states that in his judgment it was thus through necessity that the rain gauge was invented.

Various types of gauges are used in different parts of the world. In Europe and Asia gauges are of glass or porcelain, with metal guards. The rain collected is poured out and measured in a small glass graduate. The ratio of the area of the receiving surface to the area of the measuring tube must be accurately determined. Perhaps the most satisfactory rain gauge is the so-called British gauge, which has a galvanized iron funnel about 127 millimeters in diameter.

The British gauge

It is of a pattern similar to the Snowdon gauge. The Snowdon gauge probably originated with Dalton. It is constructed almost exclusively of copper or galvanized iron; the funnel is fitted with a ring of turned brass. The British rain gauge, with respect to funnel, brass ring, inner can, and glass bottle, is identical with the Snowdon gauge; the outer can of the former, however, is made of a cheap combination of lead and iron. Should the outer can become damaged or leaky, the efficiency of the gauge is not thereby impaired. This gauge has a glass bottle, a graduate glass measure, and is certified by the British Rainfall Organization, an organization founded by the late G. J. Symons. The society collects records of rainfall from nearly 5,000 voluntary observers. In the United States many different gauges have been used.

The Snowdon gauge

The Weather Bureau uses a three-part metal gauge somewhat larger than other gauges. There is a galvanized iron top-piece with brass-turned edge, and funnel shaped. This fits as a sleeve on a cylinder of galvanized iron. The diameter of the collecting surface is about 203 millimeters, and the length of the cylinder about 510 millimeters. This cylinder serves in case of overflow to retain excess rain; and is also available for snow measurements. The third piece of the gauge is a copper or brass cylinder with a diameter of about 64 millimeters. The rainfall is collected and measured in this inner tube. The area of the collecting surface is about a thousand square millimeters, or ten times the area of the inner measuring surface. The catch of rain is, therefore, magnified ten times. A wooden stick, generally cedar, is graduated in millimeters or inches, and, proper allowance being made for wetting and displacement, the depth of rain is read on the wet stick. One millimeter will represent one tenth of a millimeter of rainfall. There are many forms of automatic and self-registering gauges, some in which the weight of the water is made to record on a moving drum, as in the well-known Fergusson gauge and its various modifications. In other types, such as the tipping-bucket, a definite quantity of rain tips a small bucket, which in tipping makes electric contact and thus records at a distance. All such gauges need frequent inspection and cleaning.

U. S. Weather Bureau gauge

Fergusson gauge

To insure accuracy of the catch, a gauge should be exposed with the collecting area, if possible, level with the ground or a foot above. There should be no near obstructions and no trees or shrubbery close by from which water could drip or be blown into the gauge. Many of the Weather Bureau gauges are exposed on the roofs of tall buildings in cities and are subject to wind eddies. The catch at nearly all the principal cities is thus too low, differing by 30 to 40 per cent from the amount which would be collected by a gauge on the ground. Other inaccuracies are met unless the gauge is provided with some form of wind screen, such as that devised by Professor Nipher. There is

Placing of the gauge

still another cause of inaccuracy. In many cities the place of exposure has been changed so often by removals that the data are not comparable. As an offset, certain nonofficial records have been maintained with great fidelity, over long periods, without change of exposure or method of measurement. Records have been maintained at New Haven for a period of 128 years; at New Bedford, 103; at Albany, 94; at New York, 90; at San Francisco, 64; and at Blue Hill, 32.

Place of exposure should not be changed

Rainfall should be measured as soon as possible after the rain has ended. The best results are not obtained by measuring once in twenty-four hours. Rain is lost through evaporation; and light showers are not recorded on many self-registering gauges, the mechanism not being sufficiently sensitive. Thus the time precipitation begins and the time it ends is not, with certainty, known if dependence is placed upon records made by such instruments. At Blue Hill Observatory an instrument known as an *ombroscope*, devised by Fassig and modified by Fergusson, shows by the discoloration of prepared paper moved by clockwork the time when the first drop falls. The wind carries the smaller raindrops past the opening, and it may be that only the time of falling of the heavier drops is thus recorded. The instrument does not record the beginning or ending of fog. Nor does any known type of rain gauge give a true record if the raindrops are small and the wind high; for these fine drops are easily carried past and even whirled out of the mouth of the gauge. There is room for improvement in this direction.

The ombroscope

55. Variation of rainfall with altitude in mountainous countries. It is evident that mountains materially influence the rainfall. Various writers have shown the effect of mountain ranges in causing an uplift of the air current and subsequent condensation and precipitation. Cloud-capped summits are familiar throughout the world, and even moderate elevations are sometimes covered with clouds when the lowlands have sunshine. Particularly noticeable are the cloud caps of high mountains or snow peaks. In these altitudes the air is cooled below the temperature of saturation,

and cloud or snow banners form, streaming out from the peak in the direction from which the wind blows,—"in the teeth of the wind," as it were. Snow banners in the Sierra are eloquently described by Muir.

Cloud banners and cloud caps

In addition to these forms there are the well-known "Tablecloth" at Cape Town, and the "Helm and Bar" at Cross Fell. Perhaps the most elaborate attempt to discuss mathematically the condensation of vapor on mountain slopes is that of Pockels, who deduces the minimum elevation at which condensation may begin. Assuming the average vertical distribution of temperature and moisture for each of the four seasons, he finds that, starting from an elevation of 0 meters, the following are the approximate heights:

MINIMUM ELEVATIONS FOR CONDENSATION

Elevation in meters	Spring	Summer	Autumn	Winter
0	725	850	405	400
500	485	710	615	760
1000	855	570	600	1070
2000	920	730	1180	1100
3000	830	1060	1208	1130
4000	700	1125	1240	1100

The intensity of condensation, and presumably precipitation, is greatest, Pockels thinks, where the slope is steepest. This conclusion is not, however, in accord with facts, and there are factors other than merely those of elevation and the ascent of the air affecting the intensity of precipitation.

Factors affecting precipitation

In California the author has found a definite increase in precipitation with elevation in the Sierra Nevada in going east from the Sacramento Valley and the San Joaquin Valley. In the lowlands the rainfall is rather evenly distributed, and on the same level the distribution, as to both intensity and frequency, is comparatively uniform. There is, however, a marked difference in the amount of rainfall at stations close together but differing in elevation.

Elevation of maximum rainfall

The amount of rain increases as one goes from the floor of the valley through the foothill section and up the mountain side, reaching a maximum at a height of about 1,500 meters.

The records of the stations along the line of the railroad from Sacramento to Summit, covering a period of about forty years, show a steady increase in the quantity of rain caught by the gauges of about 75 millimeters for every 100 meters rise in elevation. The rate of increase is greatest about the 1,000-meter level, and becomes negative above the 1,500-meter level. At these high levels, however, much of the precipitation falls in the form of snow, and it is possible that with our present methods of reduction true values have not been obtained.

Rainfall east of the Sierra crest

East of the Sierra crest precipitation decreases rapidly with decrease in altitude, maintaining a constant rate to the 1,500-meter level and a diminishing rate below this elevation. The distance and precipitation curves conform to the profile in general shape, except that their maxima are west of the topographic crest, occupying the same relative position with respect to the Great Valley as the 1,500-meter level. They have a tendency to become horizontal over the level portion of the profile, to rise over western slopes below the 1,500-meter contour, to fall over western slopes above this, and to fall over eastern slopes. In other words, the general slope of the country seems to have more to do with the amount of precipitation than does altitude.

Rainfall upon the plains of Northern India

Precipitation upon the plains of Northern India and the southern slope of the Himalayas exhibits a similar variation. An empirical equation giving the relation of precipitation and elevation has been developed from observations in that region, as follows:

$$R = 1 + 1.92\,h - 0.40\,h^2 + 0.02\,h^3,$$

in which R represents the amount of rain and h the relative height in units of a thousand feet above an assumed plane, which was itself 300 meters above sea level. The critical elevation was 1,270 meters above sea level, and observations were sufficient to determine that the form of the curve above this elevation was similar to that below, the complete curve approximating a cubic parabola whose axis is the line represented by the critical elevation.

Wilhelm Krebs called the author's attention to the results

of measurements made by him on the Storm and Draken mountains in South Africa, published in the *Deutsche Rundschau für Geographie und Statistik*, August, 1890, and September, 1908. Krebs applies the same process to the Sacramento and San Joaquin records, using as the base the valley floor from Sacramento to Stockton. He finds that the ratio of Auburn elevation to Gold Run elevation is as 1 to 2.4; the precipitations as 1 to 1.6; and the gradients of the slope as 1 to 1.5. The ratio of Auburn and Blue Cañon is as 1 to 3.6; the precipitations as 1 to 2.1; and the gradients of the slope as 1 to 1.9. He also compares Mokelumne Hill, West Point, and Bear Valley. The ratio of precipitation agrees with the gradients of slope and differs from the ratio of elevations.

In Africa and in California

Probably the best rainfall record of exceptionally heavy precipitation is that made at Baguio, Philippine Islands, July 14–15, 1911. The chart (Fig. 81) taken from the *Manila Weather Bulletin* shows that the total rainfall from noon, July 14, to noon of the next day was 1,168 millimeters (45.99 inches). The greatest hourly amounts were 91 and 90 millimeters; the greatest ten-minute rainfall was 18 millimeters; and the greatest five-minute fall was 10 millimeters. The total precipitation at Baguio for the four days, July 14–17, was 2,239 millimeters (88.15 inches).

Exceptionally heavy rainfall

The heaviest monthly rainfall in the United States occurred at Helen Mine, Cal., in January, 1909, when 1,817 millimeters (71.54 inches) fell. The heaviest twenty-four-hour rainfall in the United States occurred at Altapass, Mitchell Co., N. C., when 564.4 mm. fell from 2 P.M. July 15 to 2 P.M. July 16, 1916. The heaviest short-period rainfall occurred during a cloudburst at Campo, Cal., August 12, 1891, when 292 millimeters fell, practically in eighty minutes.

Heaviest rainfalls in the United States

The table on p. 216 gives the rate and duration of the heaviest known rainfalls.

56. Measuring rainfall by rings of annual growth. Many attempts have been made to determine secular variation in rainfall by studying the rings of annual growth on trees. Manson and others have made cross-sections of the *Sequoia*

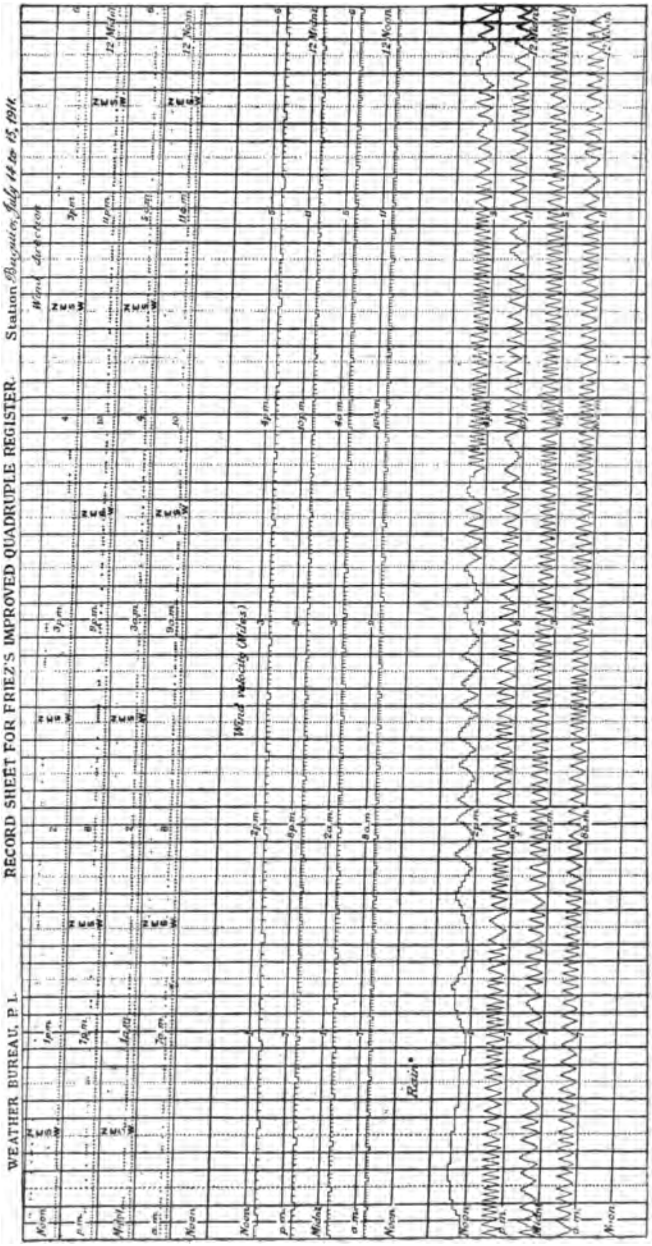


FIG. 81. HEAVIEST RECORDED RAINFALL

The recording pen traces its record in a zigzag line of steps, each step representing 0.254 mm. of rain, and one complete zigzag containing just ten steps or 2.54 mm. of rain. In 24 hours there were 1168 mm. (46 inches) of rain.

gigantea, also of the redwoods of California, and have examined the consecutive character of the rings with reference to the known rainfalls and thus tried to discover if there were any periodicity in rainfall. Huntington has also made extensive

Place	Date	Rate per hour in mm.	Actual duration of rate in hours
Baguio, P.I.	14, VII, 1911	49	24.0
Cherrapunji, Khasi Hills, India*	14, VI, 1876	43	24.0
Montell, Tex.	28, VI, 1913	28	18.5
Concord, Pa.	5, VIII, 1843	135	3.0
Campo, Cal.	12, VIII, 1891	219	1.3
Guinea, Va.	24, VIII, 1906	470	0.5
Curtea de Arges, Rumania	7, VII, 1889	615	0.3
Graz, Austria	3, VII, 1914	110	0.5
Galveston, Tex.	4, VI, 1871	429	0.23
Fort McPherson, Neb.	27, V, 1868	428	0.083
Kansas City, Mo.	6, IX, 1914	195	0.083
Valdivia, Chile	11, VI, 1912	480	0.0083
Malta	16, X, 1913	51	3.0
Amboina (Dutch Indies)	6, VI, 1912	120	0.083
Rogodjampi (Dutch Indies)	10, II, 1912	192	0.083
Patjet (Dutch Indies)	14, XII, 1912	204	0.083

*Blanford, in his *Climates and Weather of India*, 1889, pp. 77 and 265, gives 1,036 mm. at Cherrapunji as the heaviest rainfall recorded for one day in India.

studies along more general lines. A. E. Douglas, of the University of Arizona,[1] has identified a long period of tree growth (Arizona pines) with certain meteorological cycles. There are four marked maxima about the years 1400, 1560, 1710, and 1865. A period of 33.8 years fits very well after 1730. The last crest came in 1900, and this can be identified with the well-known Brückner period. There is a rather persistent period of about 21 years, its last crest occurring in 1892. This pulsation is well marked from 1400 to 1520; then for a hundred years fails or shows discrepancies, and after that, from 1610 to the present time, is marked and regular. There is also a period of 11.4 years which is practically the sunspot cycle.

Four marked maxima

[1] "Method of Estimating Rainfall by the Growth of Trees," *Am. Geog. Soc. Bull.*, May, 1914.

Correlation between tree growth and sunspot variation is not confined to the trees of Arizona and California, for measurements made on thirteen tree sections from the forest of Eberswalde indicate a relation between tree growth and rainfall, temperature, and solar conditions. Apparent climatic cycles have been investigated, and, what is of real importance, a method of estimating rainfall has been found which may be susceptible of extension and adaptation to other fields of science.

Tree growth and sunspot variation

57. Rainfall distribution. *Isohyets* are lines denoting equal amounts of rainfall, and until within a year or two they have been the only lines used in charting rainfall. *Isomers* are lines of equal proportion of rainfall; that is, lines which show the rainfall of a given period in percentages of the total rainfall. But in all rainfall maps thus far made, no correction has been applied for either elevation or temperature. The first rainfall map was published in the *Physical Atlas* of Berghaus in 1845, and showed isohyets for Europe. Loomis, in 1882, drew the first chart of world isohyets, and these were subsequently redrawn by Buchan. About 1898 Supan issued his charts. Herbertson soon followed, the latter being the first to give monthly values. In Bartholomew's *Atlas* may be found the most comprehensive series of rain maps. From such charts it may be seen that the equatorial regions in general have a precipitation amounting to about 1,000 millimeters. Eastern coasts receive a relatively heavy rainfall, especially if mountainous. The rain is heaviest where the trade or monsoon winds blow directly toward the land. At points from 20 to 35 degrees either side of the equator there is a marked decrease in rainfall, and dry zones and deserts appear west of the regions influenced by the trades and monsoons.

Isohyets

Isomers

Rainfall in equatorial regions

There is, on the whole, a steady diminution of rainfall from equator to pole corresponding with the diminution of temperature and of vapor-carrying capacity of the air. Three exceptions should be noted: (1) The coastal lands, where sudden changes of temperature are frequent and the air is nearly saturated with moisture, are rainy,

Decrease poleward

except where cold currents well up and make a cool area near the coast, as happens near the tropics on the west coasts. (2) Great temperature changes also occur in mountain lands, and where the air is sufficiently damp, rain is common. (3) The hearts of the continents, far from the source of water vapor in the oceans, and the regions reached by winds blowing out from them over dry land, are very dry. There is a great area of excessive rainfall (over 2,000 millimeters) over the Atlantic between Newfoundland and Ireland; and a more restricted area on the northwest coast of the United States, including Alaska. From Cape Flattery to Cape Mendocino the annual precipitation ranges from 2,500 to 3,000 millimeters. Southward from Cape Mendocino the rainfall decreases; near San Francisco it is about 600 millimeters; and at San Diego about 250 millimeters. Types and percentages of rainfall in the United States are shown in Fig. 82.

Areas of excessive rainfall

The term "norm" is frequently used in connection with rain to represent the amount that would be precipitated provided the rain were uniformly distributed throughout the period under consideration. Thus, if the total annual fall is divided by 365 and the quotient multiplied by the number of days in each month, we get the norm for the month. If, as is the practice with some writers, a value of 100 is given to the norm, then the amounts can be expressed in percentages. Thus, if at a given place the norm, or normal rainfall, for January is 2.5 millimeters, then for a month with 3 millimeters the corresponding norm would be 120.

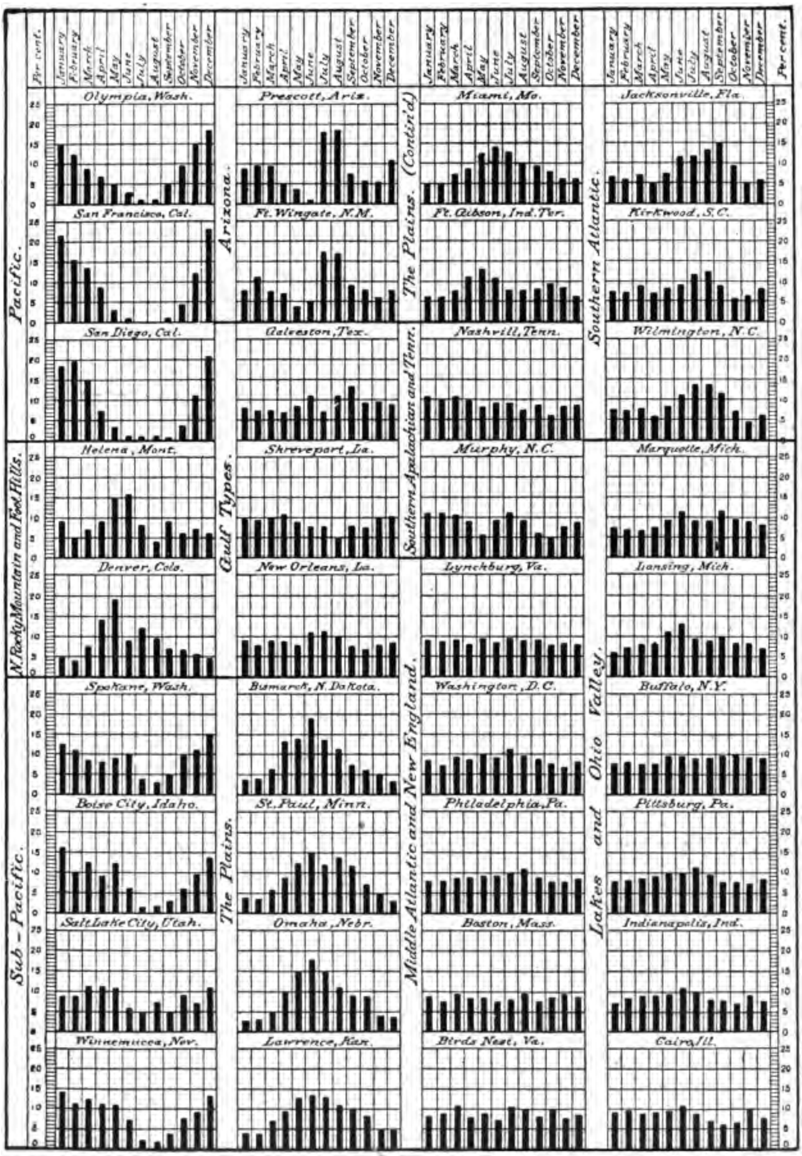
The "norm," or normal rainfall

The term "pluviometric coefficient" was introduced by Angot to indicate the ratio of the mean daily rainfall of a particular month to the mean daily rainfall of the whole year. Wallis[1] uses the term "equipluves" for the lines of equal departure from the rainfall norm. Mill uses the term "splash" to represent the distribution of rain in a shower, while the term "smear" represents the generalization of a succession of splashes. More appropriate terms are needed.

The "pluviometric coefficient"

"Splash," "smear"

[1] *Monthly Weather Review*, Jan., 1915.



From Bulletin Q, U. S. Weather Bureau

FIG. 82. TYPES OF MONTHLY DISTRIBUTION OF PRECIPITATION IN THE UNITED STATES

In addition to charts of rainfall for seasons and months, Mill, Reed, and others have attempted to chart the distribution of rainfall, using the cyclone or barometric depression as a unit. The need of some unit other than a purely arbitrary time division has been advocated by Ward, and the cyclonic unit seems suitable for many purposes in climatological work. The term "smear," which Mill introduced in his study of rainfall distribution accompanying the passage of cyclones across the British Isles, has been used by Reed in discussing cyclonic rain for certain storms crossing the United States. While the smears for British cyclones show large continuous areas for rainfall of 25-millimeters depth, such continuity is not so well marked in the smears for the United States. There does, however, seem to be a relation between areas of heavy precipitation and extensive water areas, such as the Great Lakes, the Gulf of Mexico, the Mississippi watershed, and the Atlantic. But there are many instances of areas of heavy precipitation being at some distance from the track of the cyclone.

Cyclonic unit for rainfall distribution

Charting rainfall

It should be pointed out, however, that in such work as the foregoing the storm tracks, as charted, are not necessarily the true paths of disturbances, but only approximations. Locating the area of lowest pressure is not the best way to find a storm center, if by the latter term we mean the center of air motion. In fact, the methods used for charting storms are practically those of thirty years ago. If we determine and represent the progressive movement of the storm by charting the air motion, laying emphasis on wind rather than pressure, we may get more accurate storm tracks. The following is a suggestion given by Bjerknes, in a lecture before the University of London: Chart the winds so as to show the horizontal direction of flow; then represent velocity by separate lines crossing the lines of flow at right angles, indicating the speed by comparative proximity. Such a chart will show lines of convergence and divergence, useful in studying cyclonic or anticyclonic motion. If topography is also shown, there will obviously be ascending currents where the winds blow toward high levels, and descending

currents on the leeward slopes. Comparing charts thus obtained with actual precipitation, it appears that the heaviest rainfall is where there are uprising currents, and the lightest where there are descending currents.

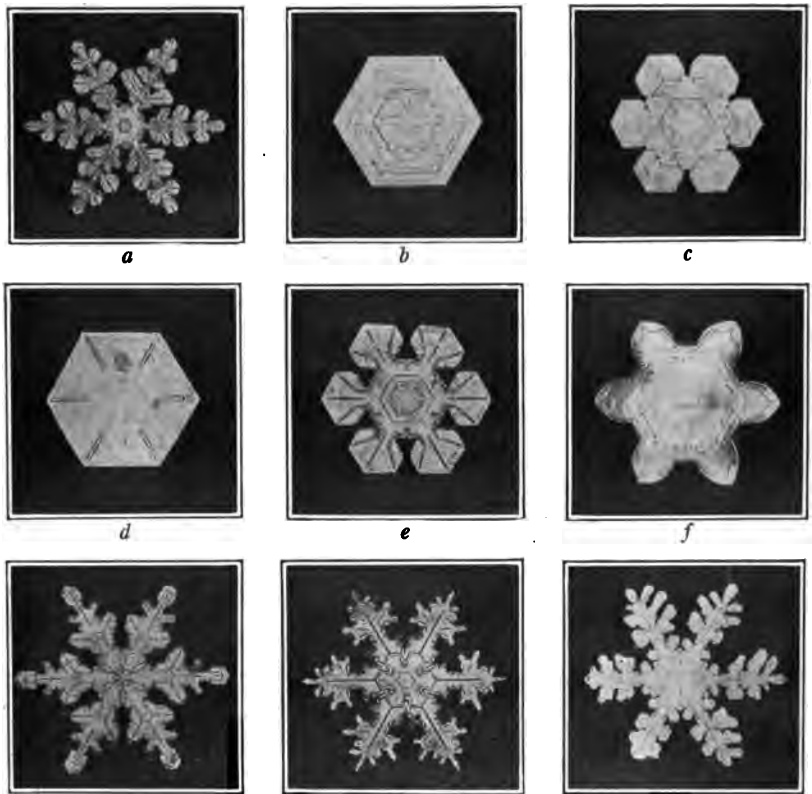
SNOW

58. Snow crystals. Perhaps the most comprehensive and at the same time most detailed study of snow crystals is that made by W. A. Bentley of Jericho, Vermont. His collection includes more than 1,000 photomicrographs of snow crystals obtained during a period of seventeen years of observation. He has classified snow crystals according to form and according to position relative to the storm center. He adopts Hellmann's general classification, dividing them into *columnar* and *tabular;* and for convenience divides these into subvarieties, using names suggested by Scoresby. Solid tabular forms are called *lamellar*, while crystals of more or less open structure possessing solid tabular nuclei resembling ferns, are called *fern stellar*. Columnar forms connecting one or more tabular crystals are called *doublets;* and extremely long, needle-shaped types, *needilar* or *spicular*.

Types of snow crystals

During cold snowfalls the solid columnar and tabular forms appear to be of nearly equal number with the more open stellar and fernlike varieties, and they considerably outnumber the granular forms. Doubtless the actual connection between forms and sizes of snow crystals and the temperature of the air is more intimate than our present knowledge would indicate, for our studies are based on air temperatures at the earth's surface instead of at the cloud levels where the crystals form. By close study of the microphotographs, Bentley finds that the most common forms outlined within the nuclear parts of the crystals are a simple star of six projecting angles, a solid hexagon, and a circle. The subsequent additions assume a bewildering variety of shapes, each of which usually differs widely from the one that preceded it and from the primitive nuclear form at its center. Examining the photograph (Fig. 83*a*), we see a crystal star as it was probably shaped at birth. In this form it was probably carried upward

Size of snow crystals and temperature of the air



g FIG. 83. h SNOW CRYSTALS i After Bentley

by ascending currents, assuming at some upper level the solid hexagonal form around the star-shaped nucleus. Becoming heavier, it probably fell, acquiring further growth in lower levels. The crystal in Fig. 83*b* probably originated at a high level and completed its growth at low levels.

Formation of crystals

The magnification in Bentley's photographs varies from 30 to 50 diameters. The author in studying the structure of snowflakes found that the most frequent form was a needle or spicula. In general the angles formed by the crossing of these needles were 60° or 120°. (Figs. 84 and 85.) The ratio of the depth of snow to water at 273° A. was 14 to 1 of the flakes in Fig. 84.

59. **Measurements of snowfall.** It may be frankly conceded that many of the measurements of snow are of doubtful accuracy. It is problematical whether the snow gauge of ordinary design, owing to clogging and wind action, catches the proper proportion of the fall. Where possible, snow should be melted and converted into water immediately upon reaching the collecting surface; but this is impracticable at most points. Snow collected in a gauge is usually brought into a warm place, melted, and then measured. Warm water may be used, if an accurate estimate of the snow equivalent is to be ascertained. In general, one inch of snow is considered the equivalent of one tenth of an inch of water, but this relation is not fixed, and the writer has found that in the same snow bank the density of the snow varies markedly. The snow sampler devised by Church, referred to below, and a form of

Water equivalent of snow

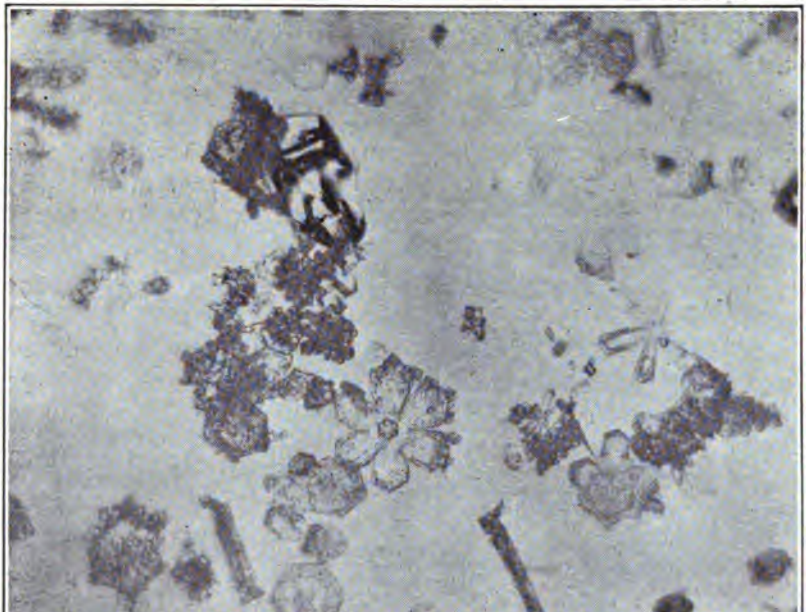


FIG. 84. SMALL SNOWFLAKES McAdie
Magnified 20 diameters.

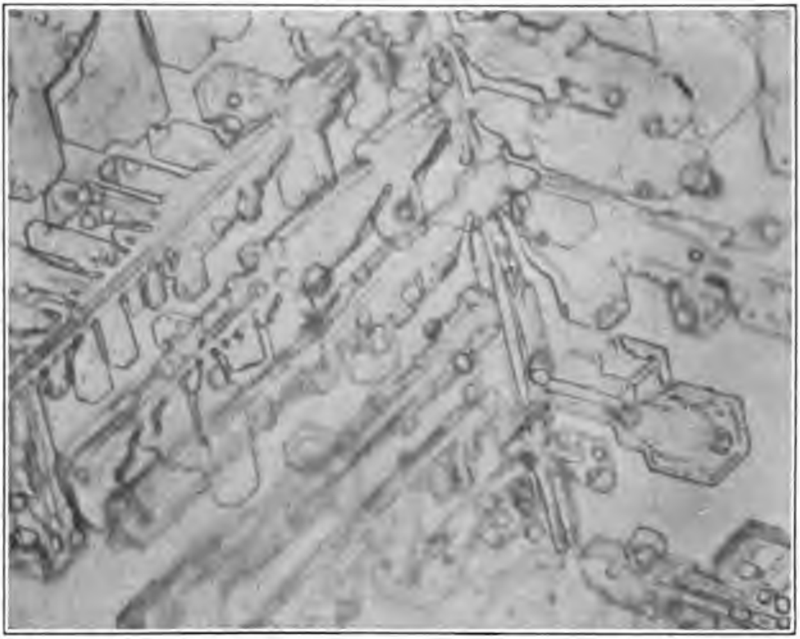


FIG. 85. STRUCTURE OF A SNOWFLAKE McAdie
Magnified 200 diameters.

density gauge, in which the given volume of snow is weighed on a spring balance, enable more accurate determination of the water content of snow than other gauges. An attempt made several years ago by the United States Weather Bureau to use snow bins was not successful.

Distribution of snow in storms

Brooks has studied the distribution of snow in two great snowstorms and finds that, in general, the snowfall is spread over a wide area on each side of the storm track; the heaviest snow comes with northeast winds and occurs in a belt about 100 to 200 miles north of the track; the northwest winds in the southwest quadrant sprinkle light snowfall over the country to a distance of about 300 miles south of the track of the cyclone center. The effects of local topography and geography make the distribution patchy.

In some of the western states, notably California, Nevada, Utah, Idaho, and Arizona, it is important to determine the probable water supply from the snow cover. For this a snow

scale, or stake, supplemented by a few density measurements, is used. Unless, however, something is known of the rate of melting and the losses through evaporation, estimates based on such fragmentary data must be taken with due allowance for error. Perhaps the most reliable data on snow measure in the United States are those made in California, Nevada, Utah, and Colorado.

Water supply from snow

For a distance of nearly forty miles the Southern Pacific Railroad has erected snowsheds in the Sierra Nevada in order to maintain traffic during the winter months. The depth of the snow is, moreover, of utmost importance to many power companies, and in mining and irrigation activities also.

60. The economic importance of snow. The problem of the conservation of snow and its dependence upon mountains and forests has been treated in detail by several writers, and more especially by Professor J. E. Church, Jr., of the University of Nevada. The water content of the snow, determined by weighing with a snow sampler devised by Church, permits quick surveys of large areas of snow, both on mountain tops and in gulches; and in due time results will be forthcoming which will enable predictions of floods or insufficient water supply, as well as studies of stream flow and water resources. The author has discussed the relation between total snowfall and depth on the ground in an effort to obtain a curve characteristic of the season. Such a curve would in effect give a measure of the heat energy expended during a given period as determined by the rate of disappearance of the snow. The disturbing factor, however, is wind action; and until we have some record of the direction and rate of flow of the air, and also of the amount of water vapor carried by the air stream, all forecasts of the probable water supply will be subject to error.

Characteristic seasonal curve

The Weather Bureau has in the spring months tried to determine the probable amount of water in the snow cover of the high levels, available for irrigation. The plan of intensive surveys in small watersheds as carried out by Thiessen in Utah gives approximate information, but of limited character. An illustration of the value of such surveys is that made in the watershed of City Creek, furnishing water to

Salt Lake City. The measurements of 1915 indicated a probable water yield of the snow cover 30 per cent less than that of the preceding season, and also that the snow condition was favorable for rapid melting and early run-off. Similar work has been attempted in the watershed of the Salt River of Arizona, above the Roosevelt Dam.

Forecasts of water supply from snow

From the records made at Summit, Placer Co., California (elevation 2,138 meters), covering a period of nearly forty years, it will be seen that the annual snowfall has exceeded 19.68 meters twice in a period of thirty-five years. In the season 1879–1880 nearly 20 meters fell. The least seasonal snowfall was in 1880–1881, the season next after that of heaviest snowfall. The total amount was less than 4 meters. The average depth of the snow is nearly 11 meters. Records of the depth of snow on the ground have been kept daily since 1898.

Snowfall records

The author has discussed[1] the method used by Professor J. N. Le Conte in 1908 for determining the mean rate of melting. It was found that the curves were extremely irregular for the eleven seasons considered up to March 1, but fairly smooth after that date. Professor Le Conte obtained an average curve showing the mean depth of snow on the ground at Summit. The curve is reproduced in Fig. 86. In many ways the period July 1, 1910, to June 30, 1911, was one of the most remarkable on record. It followed a season when there was less snow in the mountains than there had been in forty years. Up to January 9, 1911, the season was exceptionally dry and the snowfall a negligible quantity. Then there occurred a rapid change to the other extreme, and the snow fell almost steadily until January 27, when there was approximately 5 meters on the ground. February was a month of moderate snowfall, also of moderate melting. The total decrease was only 330 millimeters. During the first week in March, 2,286 millimeters of snow were added. Before the middle of the month a total depth of 7,740 millimeters was recorded. Then followed a long period of fair weather, permitting a rapid and nearly uniform rate of melting, the depth decreasing at

[1] In *Monthly Weather Review*, June, 1910.

the rate of about 203 millimeters a day. It was noticeable, though, that the melting was less rapid as the depth

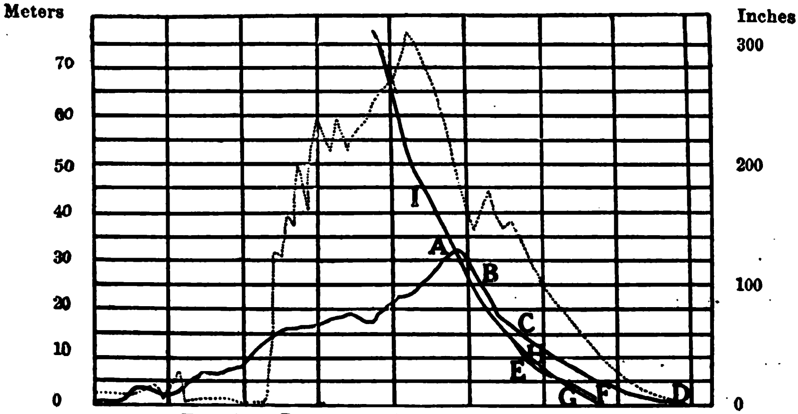


FIG. 86. DEPTH OF SNOW AT SUMMIT, CAL.
Solid line indicates average depth of snow, mean of ten seasons. Dotted line, depth of snow, 1910–1911. *ABEFG*, average rate of melting.

decreased, although with the natural increase in the length of the day and the approach of warmer weather the contrary might have been expected. Without doubt, the packing process plays an important part, and all measurements of depth and of melting must be corrected for this factor. The author once made some approximate measurements of the water content of snow as an experiment bearing on this problem. Samples of snow were taken from the top and the bottom of a snow bank which had a depth of about 3.6 meters. Melting the samples, it was found that it required about 508 millimeters of the loosely packed snow at the top to make 30 millimeters of water, while of the more compact, almost slushy, snow at the bottom it required only 102 millimeters to make 25 millimeters of water.

In the diagram (Fig. 86) the dotted lines show the depth of snow on the ground for the season 1910–1911. The solid line represents the mean depth of snow, and the curve marked *ABEFG* is the mean rate of melting from March 1 to May 26, as determined by Le Conte.

The author has also designed a means of comparing the actual curve of melting for any given season with the mean

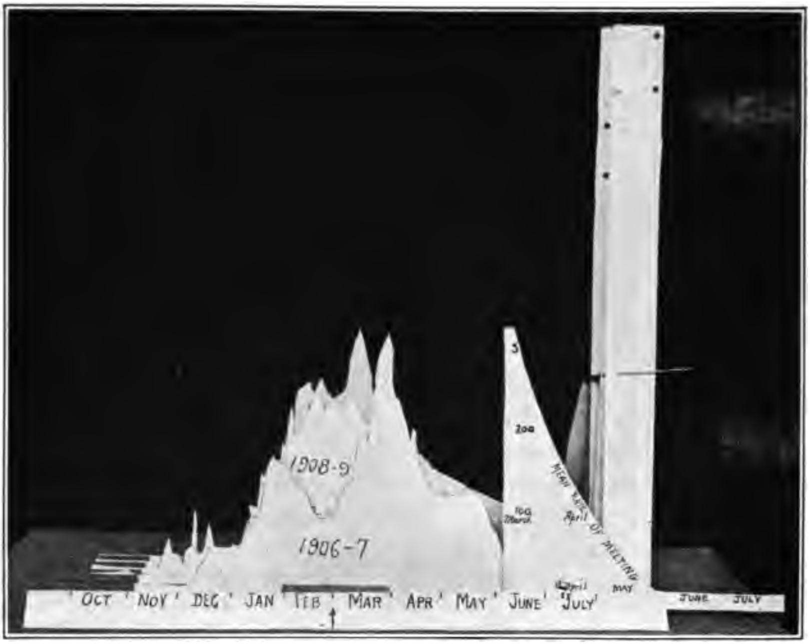


FIG. 87. METHOD OF STUDYING SNOW COVER IN THE MOUNTAINS AND PROBABLE RUN-OFF

curve, so that one can determine the probable date of the disappearance of the snow. Fig. 87 represents a wooden base with reinforced pieces at suitable intervals. Small grooves are cut in the base plate, and in these are inserted pieces of bristol or cardboard cut to represent the depth of the snow. As there is practically no snowfall of importance during July, August, and September, the year begins on October 1. March 1 falls about the middle of the design, and we thus have on the left the snowfall of winter, while on the right side we have the snowfall of the spring months. In the first groove there is inserted a card showing the mean rate of melting. This can be slid along in the groove and the rate compared with that of any given year by bringing into line the two profiles. While this method cannot, in strictness, be said to give the true rate of melting for the whole year, it affords an approximate measure of the rate of melting under normal conditions. Snowstorms are

Means of computing rate of melting

shown by peaks which indicate both the added depth and the rate of melting in the intervals of fair weather. At the right side of the frame vertical strips of cardboard show the total precipitation for the season, and, by means of suitable notches, how much of the precipitation was snow. The whole design enables one to compare readily the depth of snow on the ground at any given date with the amount during previous seasons.

Summit is an interesting station for snowfall work, because 86 per cent of the precipitation falls in the form of snow, and most of the rain falls in July, August, and September, practically before the snow cover amounts to anything. It is true that occasionally there will come a warm rain in January or February, and such a condition rapidly reduces the depth of snow. The greatest factor, however, in reducing the depth is probably the wind. Under certain conditions, when the dry air moves rapidly from the northeast, decrease by evaporation becomes excessive. Two or three such days will lower the depth 200 millimeters or more.

Wind an important factor in melting

In a general discussion of the snowfall of the eastern United States, C. F. Brooks has shown[1] that on account of low temperature and dampness the Lake region, Appalachians, and North Atlantic coast get the heaviest snowfall. The Ohio Valley, South Atlantic States, and Gulf States are usually too warm for much snow. In the Northwest the snowfall is moderate because of the winter dryness. Within the larger provinces snowfall is locally modified by topography and exposure to moist winds. Thus the Appalachians get heavier snows on their western than on their eastern slopes (except in Vermont), and the eastern shores of the Great Lakes get more snow than the western.

Regions of heaviest snowfall

Snow generally falls in connection with winter cyclones, because the cyclonic action and effect of topography cause precipitation. The northeast wind is the wind of great snowstorms. The northwest wind, though cold, is generally dry and so brings at most only snow flurries, except locally on a windward mountain slope or in the lee of the Great Lakes.

Greatest snowstorms from northeast

[1] In *Monthly Weather Review*, Jan., 1915.

ICE STORMS

61. The various types of ice storms. Bonacina has discussed the general problem of frozen precipitation in *Symons's Meteorological Magazine*, confirming the theory of Hellmann,[1] Dansey, and others that rain falling in a temperature below the freezing point produces the phenomenon of "glazed frost," or "silver thaw"; and this is due to the presence of a warm stratum above the cold surface air, and is not in any way an instance of supercooling, or, more properly, subcooling. "In England," Bonacina says, "a warm southerly current will often climb over the shoulders, as it were, of a cold, easterly surface current, ultimately replacing it; and in such cases the premonitory symptom of a thaw is liquid rain or ice-rain, falling while the surface temperature is still well below the freezing point. In this manner we are robbed of many an expected snowstorm. The cold surface air tends to freeze the rain, and so, instead of liquid rain, we get ice-rain. And this leads us to recognize five distinct species of frozen precipitation. These, excluding rime or hoarfrost, which is of the nature of deposition rather than precipitation, are snow, hail, graupel, sleet, and ice-rain.

"Glazed frost"

Snow, hail, graupel, sleet, ice-rain

[1] An elaborate exposition of the various forms of water vapor is given by Hellmann in the *Königl. Preuss. Met. Inst.*, 1915, translated by C. Abbe, Jr., in the *Monthly Weather Review* for July, 1916. Hydrometeors, in the narrow sense of the word, are only those forms of condensation that bring directly to the earth water in its liquid or solid form, so that the clouds (forming a chapter by themselves) are not included. The forms are:

1. Direct condensation at or near the earth's surface:

Liquid		*Solid*	
"Sweat"	(Beschlag)	Frost	(Reif)
Dew	(Tau)	Mist ice	(Nebeleis)
Mist	(Nebelwasser)	Ice fog	(Eisnebel)
Wet fog		Rime	Rauhreif / Rauheis
"Scotch" mist	(Nebelreissen)		
Fog drip	(Nebeltraufe)		
Rain without clouds		Snow without clouds	

2. Direct condensation in the free air:

Water clouds		Ice clouds	

3. Indirect condensation in the free air:

Rain	(Regen)	Snow	(Schnee)
		Graupel	(Graupeln)
		Hail	(Hagel)
		Sleet	(Eiskörner)
		Glaze or glazed frost (Glatteis)	

"Snow is by far the most important in the economy of nature and is produced by the direct passage of aqueous vapor into the frozen state; it is, indeed, next to rain, the most important of all forms of atmospheric precipitation. Hail (hard or true hail) is apparently a product of thunderstorm activity, and this, together with its peculiar alternate structure, leads one to suppose that the freezing raindrops are carried up and down by currents many times before they finally strike the ground. Graupel (soft hail) is the little white pellets so frequent in moderately cold weather; as it does not occur in severe cold, it seems reasonable to ascribe its origin to the passage of aqueous vapor first into liquid drops which freeze before falling. Sleet is the well-known mixture of raindrops and snowflakes. Finally we have the form referred to above as being associated with glazed frost and as being symptomatic of thaws. It takes the form of plain pellets of colorless ice, and being the result of the direct freezing of falling raindrops may be called simply ice-rain."

62. Air temperature, rain temperature, and temperature of objects. C. F. Brooks has discussed at some length the ice storms of New England. An ice storm (*verglas; glatteis*) occurs when raindrops falling on trees and other objects cover them with ice. Using the Blue Hill records for various air levels, Brooks shows that there are a number of combinations of different conditions of air temperature, rain temperature, and temperature of the object relative to freezing which may produce ice storms. These are:

Conditions producing verglas

I. Temperature of the air below 273°A.
II. Temperature of the air above 273°A.
  A. Temperature of the rain below 273°A.:
    1. From passing through a stratum of cold air;
    2. From cooling by evaporation in non-saturated air.
  B. Temperature of the rain above 273°A.
  Temperature of the object below 273°A.:
    1. From residual cold;
    2. From cooling by evaporation in non-saturated air.

As is readily seen, no heavy ice storm can take place with the surface-air temperature above 273°A. In fact, no considerable ice storm occurring under this condition has been

noted at Blue Hill. However, from theoretical considerations they are possible. Raindrops may be cooled far below 273°A. without solidifying. It is well known that fog particles remain in the liquid state at temperatures far below 273°A. The lowest air temperature recorded at Blue Hill while rain was falling was 260°A. Undercooled raindrops freeze almost instantly if they strike one another or an object. Sometimes, when several such drops come together, large pieces of ice may be formed. For instance, during the heavy ice storm of February 26, 1912, in Cambridge, Mass. (air temperature about 271°A.), the diameters of the raindrops averaged about 0.5 millimeter; but the smallest frozen raindrops were 1.5 millimeters in diameter and the largest spherical ones 4.5 millimeters, while some rice-shaped pieces of ice were 6 millimeters long.[1]

Liquid rain below the freezing point

Sizes of frozen raindrops

For a few minutes during an ice storm on January 31, 1914, some sleet fell. The individual pieces were generally spherical, but many were angular; some were flat on one side, and others were agglomerations. The frozen drops varied between 0.5 millimeter and 4 millimeters in diameter. The structure of the sleet showed plainly the general conditions of air stratification. Snowflakes had passed into an air stratum whose temperature was above 273°A.; and some of them, reaching the colder layer below before entirely melting, froze. This is shown by the presence of snowflake skeletons, angularity, and the almost invariable presence in the ice of minute bubbles. The unfrozen rain which reached the ground was rain formed within the warmer stratum and of entirely melted snowflakes. The short duration of the sleet may be explained by assuming that, before and after its occurrence, the warmer layer was so thick that none of the snow passed out of it partially melted.

If the temperature of the object is above 273°A. the ice will partially melt and fall off entirely, or else be frozen,

[1] A short article entitled "The Crystallization of Undercooled Water," by Borus Weinberg, may be found in the *Monthly Weather Review*, 1909, pp. 14–15. See also W. Meinardus in *Das Wetter*, Nov., 1898, p. 247, the ice-rain of Oct. 20, 1898; T. Okada in *Jour. of the Met. Soc. of Japan*, May, 1914; W. H. Dines in *Symons's Meteorological Magazine*, Dec., 1913; E. Gold, *ibid.*, Jan., 1914.

icicles forming from melted water. When the temperature of the air is above 273°A. the ice coating will be continually melting, but no icicles will form.

When the temperature of the raindrops is above 273°A., as it generally is, a smooth coating of ice forms; also icicles from the water which did not freeze before running off. The windmill type of anemometer then becomes a fine, glistening, star-shaped affair. A temporary ice storm may occur when both the rain and the air are above 273°A., the temperature of the object being below that point.

The ice formed in ice storms is generally smooth, but it may be temporarily rough owing to an admixture of snow or sleet.[1] Another kind of roughness may develop if the water freezes slowly after it has fallen,— ice crystals a foot or more in length may sometimes be seen forming on a cement sidewalk during an ice storm.

Upper-air condition in ice storms

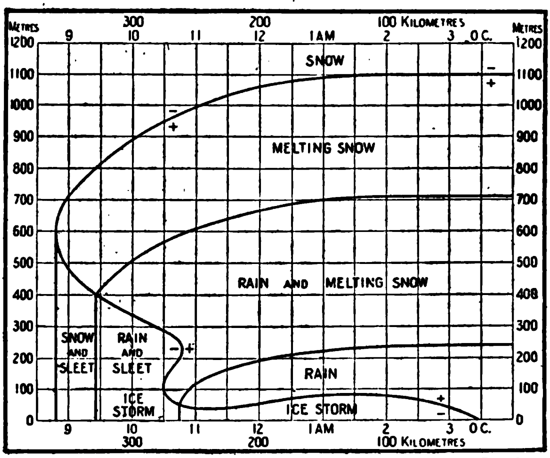
Upper-air conditions of temperature now require attention after the foregoing discussion of the effect produced. To show the mode of formation of the various kinds of precipitation, the accompanying diagram (Fig. 88) has been constructed from Blue Hill observations of January 5–6, 1910. It shows the probable conditions of precipitation and temperature from a height of 1,300 meters to the ground at sea level. The time of arrival of each stage at Blue Hill is indicated at the foot of the diagram. Fig. 89 shows the results due to the passage of these conditions at the summit and valley stations. From Fig. 88 it may be seen that the ice storm lasted about six hours at the valley station; a little over an hour at 195 meters above sea level (summit), and that a station above 400 meters' altitude would have experienced no ice storm at all. Thus local topography has a great effect on the intensity and extent of an ice storm.

[1] In Blue Hill practice "sleet" is any frozen form of falling precipitation which is not snow or summer hail. The Weather Bureau has decided to restrict the term "sleet" for official purposes to the small particles of clear ice which frequently fall in winter with or without an admixture of rain. The term "rime" is used for a coating of rough ice formed on terrestrial objects from fog when the water droplets are undercooled so that they turn to ice on coming in contact with solid bodies. Rime is what the French call *givre* and the Germans *rauhreif*.

63. Wind conditions which produce ice storms. What pressure and wind distributions cause these ice-storm inversions of temperature? There are three general wind conditions which produce ice storms:

1. Warm air arriving over residual cold air.
2. Cold air coming in below and warm air arriving above.
3. Cold air pushing in from the north or west below a rain cloud.

The ideal conditions for the first are, that after the strong radiation of heat from the lower-air strata to the ground in an anticyclone, a cyclone rapidly advances toward New England. The storm represented in Figs. 88 and 89 is an excellent illustration of this type (which will be called the "southerly" type). Within twenty-four hours a strong anticyclone (1,043 kilobars) over New England had been replaced by a cyclone from the west-southwest, well supplied with warm moist air by the south winds in front of a trough of low pressure extending to the Gulf of Mexico.[1] The record in Fig. 89 shows a regular rise in temperature at the summit, but an irregular rise in the valley. Special notice should be taken of the change between 3 A.M. and 4 A.M. January 6, at the lower level. This clearly illustrates local air drainage.



Brooks

FIG. 88. CHART OF CONDITIONS DURING ICE STORM JANUARY 5–6, 1910

The ideal conditions for the second, or "northeasterly," type are, that while an active cyclone in the south is supplying plenty of warm air, there is an anticyclone in the north

[1] For a theoretical discussion of the warm south wind setting in over cold stagnant air, see W. Schmidt, "Weitere Versuche über Böenvorgange und das Wegschaffen der Boden Inversion," *Meteorologische Zeitschrift*, Sept., 1913, pp. 446–447.

giving cold air. The northeast wind blowing toward the southern cyclone brings in the cold air from the anticyclone, and the warm south wind of the east part of the cyclone brings in the warm air above. The undercurrent of cold air is often not cold enough to counteract the warming of the southerly wind, and so the temperature may rise slowly instead of falling or remaining stationary.[1]

An example of this northeasterly type of ice storm is that of February 19–22, 1898. During this time the temperature at the top of the hill varied but one degree (272° A. to 273°A.); the valley temperature ranged between 274°A. and 277°A. Thus the storm was confined for the most part to Blue Hill. The wind was constantly from the

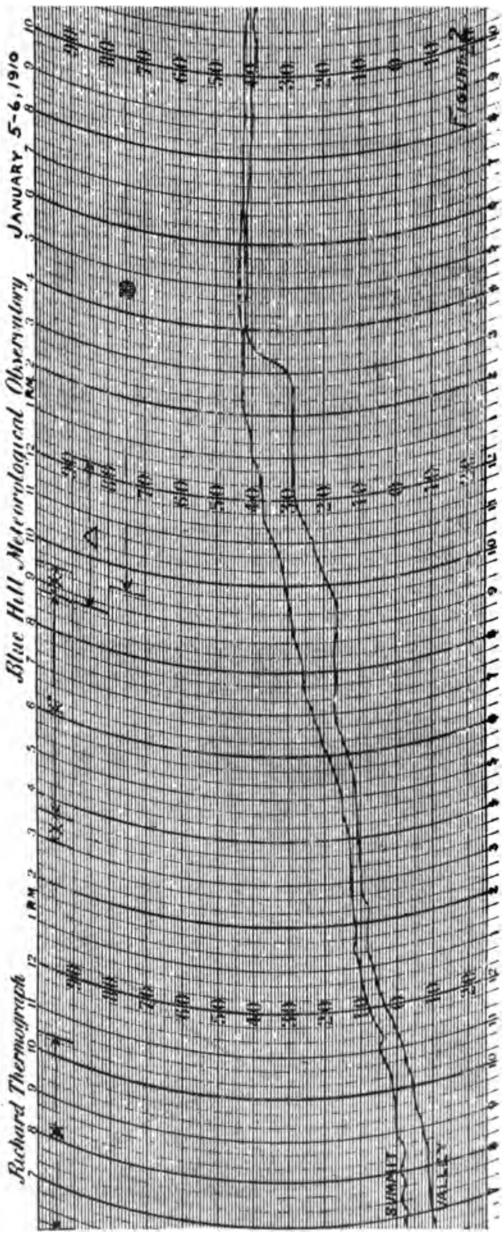


FIG. 89. TEMPERATURE RECORD DURING ICE STORM

[1] C. F. Brooks, "Three Ice Storms," *Science*, Aug. 8, 1913, pp. 193–194.

northeast and its velocity was 5 to 15 meters per second; 69 millimeters of rain fell, and 13 millimeters more of sleet. A low-pressure area was deadlocked south of a large high.

Another storm of this type was that of December 23, 1908. The temperature was 261°A., and the wind north by east, 10 meters per second. This ice storm occurred during a short interval (9:10–10:55 A.M.) in which the light snow changed to rain. An anticyclone was settling over New England and a cyclone was approaching from the Gulf of Mexico.

"Northeasterly" storm during kite flight

A third storm of this type is worth mentioning, for it occurred during a kite flight, February 9, 1905. Throughout the kite flight the wind was east-southeast, but the temperature was gradually falling (272°A. to 271°A.). The valley temperature was a few degrees higher. Up to a height of nearly 800 meters the vertical decrease of temperature was nearly the normal adiabatic; but at 885 meters there was an inversion to 273.5°A. from a minimum of 270°A. at 760 meters. At that level was the base of an arriving warm southeast wind, all the precipitation up to an hour and a half previously having been snow. On account of ice accumulations on the lower kites and wire, a kite flight during an ice storm is never long continued, nor are high altitudes reached.

Two types of ice storm over the same area at different levels

A storm of this type, which occurred over all north Germany, October 19–21, 1898, is fully described[1] by Meinardus. This storm was well observed as to conditions up to a height of 2,500 meters. On the morning of the 20th, in the lower levels, a poorly developed cyclone was centered in southwest Germany, causing a northeast indraft of cold air from an anticyclone over west Russia. At an altitude of 2,500 meters the center of the storm was over central Germany, and it was well developed. At that altitude, it caused over eastern Germany a warm south wind, six to nine degrees warmer than the northeast wind below. Over western Germany there was a cold northwest wind over the warmer northeast wind. A mountain station, 1,600 meters high, experienced this and had an ice storm from the rain that fell from a warm layer

[1] *Das Wetter*, Nov., 1898, pp. 247–260.

above it, and there was also an ice storm below from the still unfrozen rain that fell through this very cold wind. Thus there were both the northeasterly and northwesterly types of ice storms in progress at the same time at different levels.

This third or "northwesterly" type is about the reverse of the first. Here the cold air apparently enters like a wedge below, while it is still raining above. The changes in the form of the precipitation which occur are in the reverse order of those of the southerly type. This process of change from rain to snow can be seen easily by moving the conditions represented in Fig. 88 from left to right over an imaginary piece of country. Under such circumstances a wedge of cold air from the northwest is represented entering below a warmer southerly or northeasterly wind. The boundary between these two masses of air is known as the wind-shift line. The passage of a wind-shift line is a common phenomenon, but it is only occasionally that conditions of temperature and rainfall are such as to make an ice storm.

The "northwesterly" type of storm

A representative storm of this type occurred February 15, 1906. At 2 A.M., with a northeast wind and rain, the ice storm (type two at the start) began, when the temperature fell to 273°A. The temperature in the valley began to fall rapidly at 2:10, and on the hill fifteen minutes later, the wind shifting to the north and increasing from 9 to 11 meters per second. Snow began, at first mixed with rain. When the rain ceased the snow continued to 3:45, the temperature having reached 266°A. at Blue Hill.

Example of northwesterly type

Owing to the fact that often a single ice storm exhibits the characteristics of two or even three of the types described above, it is necessary to classify them according to the positions and movements of the cyclones and anticyclones which produced the ice-storm conditions. Thus two large divisions have been made, more or less arbitrarily. The first includes those storms in which there were anticyclones in the north dominating southern cyclones; the second includes those in which the cyclones and anticyclones were in regular sequence.

All ice storms of two large classes

It is noteworthy that all the ice storms occurring when a northern high dominated a southern low were of either type two or of types two and three combined. Eleven out of the thirty-one occurring under these conditions were severe.

Most of the ice storms occur when the cyclones follow the anticyclones from the west or southwest, severe storms being most common when the cyclone comes from the southwest (Gulf of Mexico). In greatest frequency come the ice storms of the northeasterly type, occurring in 116 instances. Next is the southerly type, occurring 67 times, and last is the northwesterly type, occurring 59 times. The northeasterly type is favored by southern lows and northern highs; the southerly type by the low crowding in close behind the high, and the northwesterly type coming most frequently when the high arrives close behind the low.

Distribution of ice storms by months

In the distribution by months in which ice storms occur, 48 came in January, 46 in February, 40 in March, 27 in December, 10 in November, and 7 in April. The average is 12 a year. The earliest ice storm in the fall came November 8–10, 1894, and the latest in the spring, April 30, 1909.[1]

Various combinations of conditions in ice storms

A study of some of the details of the ice storms shows that they may occur with a temperature as low as 260°A.; that it may rain hard or lightly; that the wind may come from any direction and blow a gale or not at all; that the temperature may rise, fall, or remain stationary; and that through the various combinations of these conditions, a single ice storm may change through all the types and then back again or stay consistently of a single type. In fact, as has been stated, provided it is raining and the air temperature is below freezing, an ice storm can occur under practically any consistent combination of meteorological conditions.

A few extraordinary features occurring during some ice storms not already mentioned may here be described. During the storm of January 22–23, 1904, the sudden changes in wind and temperature were the most remarkable ever

[1] The unpublished details of the 178 ice storms to March 25, 1914, have been tabulated and placed in the library of the Blue Hill Observatory.

recorded at Blue Hill Observatory. Fig. 90 is a tracing, for comparison, of the thermograph curves of the valley and summit stations on the same sheet; it shows a marked inversion of temperature, and also the variations in temperature due to warm south winds at the upper level, which did not affect the lower stratum. It also illustrates the relation of gusty winds and temperature. There is an extraordinary inversion of 8 degrees indicated, which lasted many hours. The sudden alternating gusts of wind seemed to affect the top of the hill only, and made the temperature go up and down 5 degrees or more, in almost as many seconds. The summit of Blue Hill was evidently near the level of the bounding surface

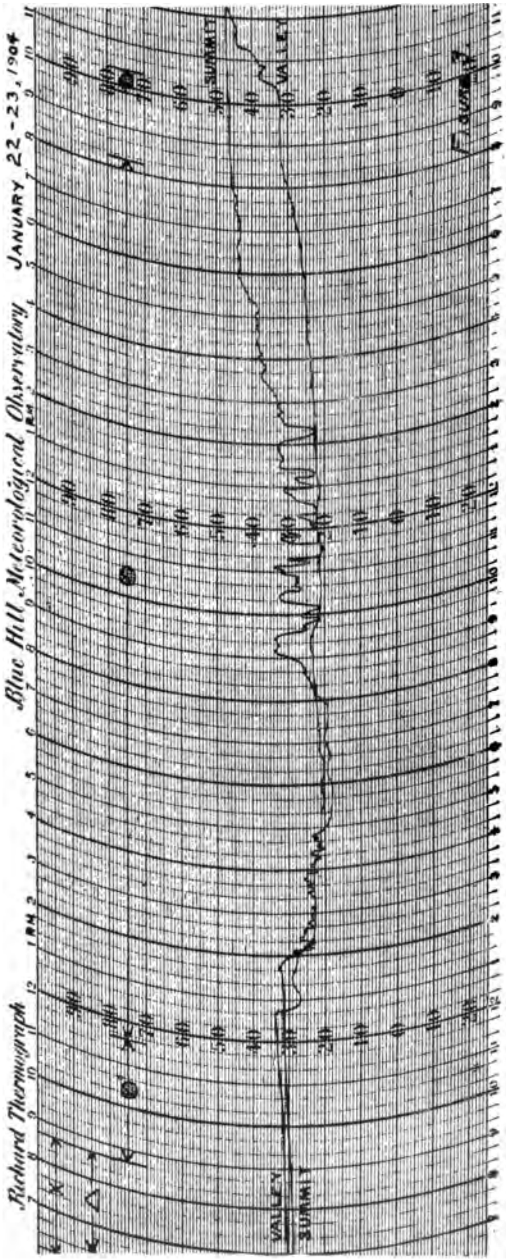


FIG. 90. TEMPERATURE RECORD DURING ICE STORM

between a warm southerly current above and a cold northerly one below. While the hill was in the upper current the temperature was high, but while the dividing surface remained above the hilltop the summit was swept by a cold north wind. In this case the line was so sharp, or its vertical movement so rapid, that the change from one current to the other was almost instantaneous. The initial cold indraft was rather significant, for it made the temperature at the valley station fall more than an hour before the temperature on the hill was in the least affected. These conditions show how slowly a warm current of air affects a cold lower stratum.

Wells

FIG. 91. ICE STORM, JANUARY 18, 1909

Figs. 91–94 illustrate the ice formation during such storms as have been described, from photographs by L. A. Wells.

Rate of cooling of raindrops

Okada has studied the rate of cooling of drops of rain falling through a cold layer. Since the rate of evaporation is the same as it is at the surface, and the radius decreases at a constant rate, which rate he found to be 0.000033 millimeter per second, the rate of evaporation was

$$q=3.3\times 10^{-6}\times 1^{2} \text{ mm. per second.}$$

Assuming the temperature of the air stratum through which the drop falls as 271.5°A., and the diameter of the drop as 2 millimeters, and the rate of fall 6 meters per second, then it would take the drop about 145 seconds to fall through the

stratum of 870 meters, which Okada found was the height at which strong winds from the southwest were blowing and rain was falling. But the rate of evaporation is proportional to the saturation deficit of the air, in this case 0.4 millimeter, so that the rate for the water drop was

Rate of evaporation

$$q \times \frac{0.4}{3.0}$$

Again, the rate of evaporation increases as the square root of the wind movement. Assume an air movement of 0.1 millimeter per second, then the rate of evaporation would be

$$q \times \frac{0.4}{3.0} \times \sqrt{\frac{1+6}{1+0.1}} = q \times \tfrac{1}{3} = 3.3 \times 10^{-6} \times \tfrac{1}{3} = 1.1 \times 10^{-6} \text{ mm/s};$$

therefore

$$t = 145 \text{ seconds};$$
$$q = 1.1 \times 10^{-6} \text{ mm/s};$$
$$\rho \text{ (density)} = 1;$$
$$\text{specific heat } c = 1;$$
$$\text{radius } r = 0.1 \text{ cm.};$$
$$l \text{ (latent heat of vaporization)} = 600 \text{ gram calories.}$$

The final temperature of the drop will be 270° A., from which it appears that raindrops falling through ice-cold layers may be sufficiently cooled by evaporation and conduction to below the freezing point and so cover the objects on which they fall with a coating of ice. In some cases the air may

FIG. 92. ICE STORM, BLUE HILL, JANUARY 18, 1909

be so moist that the drops would cool only to the dew point after falling a few meters from the mother cloud. In that case, condensation, instead of evaporation, would begin on the drop surface and there could be no glazed frost.

FIG. 93. ICE STORM, FEBRUARY 18, 1910 Wells

The theory of the formation of hail is given in the chapter on thunder storms (pp. 184–188). Many reliable records of hailstones weighing more than 150 grams can be found in Rollo Russell's book *Hail*, published in 1893. For many years it has been the practice in certain parts of Europe to attempt to prevent the formation of hail by shooting vortex smoke rings from cannon of a certain make. Despite general assurances made by those interested in the sale of such dissipators, careful and disinterested investigation fails to support the contention that such means are of any real value. They are in the same class with the attempts made in the United States, Australia, and elsewhere to produce rain by bombardment or to facilitate condensation and precipitation by explosions.

DEW, HOARFROST, GIVRE

64. Formation of dew and frost. Dew (*rosée*) (*tau*) and hoarfrost (*gelée blanche*) (*reif*) differ in their manner of formation solely in the conditions of temperature under which they are produced.

Dew is the name given to the drops of liquid water

condensed on objects from the atmosphere. The precipitation is due to the cooling of the objects by radiation below the dew point of the atmosphere. It is for this reason that dew is most frequently observed on fine evenings or nights, on the horizontal surfaces of objects possessing small capacity and conductivity for heat (provided they are insulated from conduction of heat from below).[1]

Formation of dew

If saturation is reached under similar conditions, but at temperatures below the freezing point, hoarfrost is formed; that is to say, minute needle-shaped crystals of white ice appear on the exposed surfaces, giving them a dull silvery appearance. Hoarfrost is not always frozen dew. If the dew point is 273°A. it may happen that dew and hoarfrost will form side by side according to the radiative power of the object. In such conditions the hoarfrost is more copious because the vapor pressure over ice is lower than over water and therefore condensation is more rapid.

Formation of hoarfrost

From the conditions attending their formation, both dew and hoarfrost occur with falling temperature.

"Givre" (rime) is a form of frost which occurs in nature as the result of two distinct processes. In view of this fact, and because that which is precipitated is in each case unlike

Wells

FIG. 94. ICE STORM, JANUARY 15, 1912

[1] For a remarkable collection of illustrations of frost crystals, showing especially hoarfrost crystals, see Wilson A. Bentley in *Monthly Weather Review*, Aug., Sept., and Oct., 1907.

in external appearance, the two phenomena should be distinguished as follows:

There are two kinds of givre.

Givre which condenses on objects

1. Givre which condenses on the objects themselves. It has the appearance of needle-shaped crystals of ice and forms when frost gives way suddenly to warm and moist weather, on objects of which the temperature is still below the freezing point. These objects must have great capacity and small conductivity for heat. Givre of this kind closely resembles hoarfrost in external appearance; but it forms, as a rule, with an overcast sky and always during a rise of temperature.

Givre deposited from the air

2. Givre deposited from the air. This is a deposit from the air of subcooled droplets of water, which congeal as soon as they come in contact with solid objects. Givre of this kind is deposited most copiously on the side of objects exposed to the wind. It forms a semi-transparent, rough covering of ice.

The English name for the coating of ice so generally known as "sleet" but erroneously so, is *glazed frost*, *verglas* in French, and *glatteis* in German. Perhaps the most satisfactory name for the coating of ice which occurs when cold rain falls on colder surfaces is *glaze*.

65. Dew deposits. In 1814 W. C. Wells published his now well-known theory of dew; and while there had been considerable experimentation on the subject before this treatise, like Howard's classification of the clouds it was so generally accepted that comparatively little experimentation followed. However, there is room for further investigation, particularly in connection with accurate determination of nocturnal radiation. Dew measurements have been practically overlooked by official weather services, and one searches in vain for records of the daily amounts. Dines[1] has shown that

Annual deposit of dew

the general belief that the annual dew deposit in England amounted to about 120 millimeters (as given in a footnote in the latest edition of Wells' *Essay on Dew*) is an overestimate. He estimates the average annual deposit of dew upon the surface of the earth to

[1] *Quart. Jour. of the Royal Met. Soc.*, July, 1873, p. 157.

be less than 37 millimeters. It is not an easy matter to record dew. Results cannot be obtained on nights when rain is falling, nor when there is much wind and rapid evaporation. And the observer must be up before sunrise at all seasons; that is, before rapid evaporation begins. Dew cannot be measured, like rainfall, at given hours. Dines found that only on rare occasions did the amount of dew exceed 0.25 millimeter. Out of 198 observations this amount was exceeded only three times. The average amount was about .001 millimeter. These amounts were determined by weighing.

Measurement of dew deposit

The *drosometer*, or dew measurer, is an instrument devised by S. Skinner. It consists of a hemispherical vacuum glass vessel, jacketed, of the type designed by Dewar for holding liquid air. The cup has a diameter of 11.2 centimeters, exposing a virtual surface aperture of 98 square centimeters. The vacuum is a good nonconductor, and the heat lost by radiation from the inner surface of the cup must be supplied from the air in the cup; as soon as this falls to the dew point, moisture is deposited. From measurements made on 34 nights, Skinner is of opinion that the annual dew deposit is about 2.33 centimeters a year. Skinner also discusses the value of the rain gauge as a dew collector. There were seven nights, for example, on which the dew exceeded 2.5 millimeters. But there was no trace of rain in the rain gauge, which was of the old Howard pattern, that is, 125-millimeter copper funnel standing in a bottle. The dew had evidently formed on the outer or under side of the metal funnel and run down outside. In general, dew forms on the upper side of a blade of grass or in a large drop at the tip of the blade.

CHAPTER XVI

Floods and Notable Storms

66. The relation between storm frequency and floods. Unusually heavy rains are nearly always followed by heavy run-off and floods. In some countries, as for example China, there is a direct relation between the duration of the winds from the Pacific, and floods. In the United States the precipitation that directly causes floods in the Mississippi can be traced to storms moving from the southwest. One of the important functions of the Weather Bureau is the forecasting of such conditions and the issuing of warnings.

Forecasting floods

These warnings may anticipate flood conditions by periods ranging from three days to four weeks. It is stated by Frankenfield in his *Report on the Floods of the Ohio and Mississippi*, in 1912, that the variations of the actual from the forecast stages in all except the precipitous mountain streams were practically negligible. For example, the warnings for New Orleans issued by the forecaster (Dr. I. M. Cline) nearly five weeks in advance were not materially changed except as to the date of occurrence of the crest stage, since numerous crevasses at times interfered with the expected crest progress.

Accuracy of flood warnings

The great drainage basin of the Mississippi River covers an area of about 3,211,729 square kilometers, or 1,240,050 square miles, about 41 per cent of the total area of the United States, Alaska excluded. There are six distinguishable grand basins, five of them watersheds of the principal large rivers, namely, the Missouri, the Ohio, the Arkansas, the upper Mississippi, and the Red (see chart showing drainage basin of the Mississippi, Fig. 95.)

Notable floods have occurred in the lower Mississippi in 1815, 1828, 1844, 1849, 1850, 1851, 1858, 1859, 1862, 1865, 1881, 1883, 1892, 1903, 1909, and 1912, the last being the greatest flood on record. A severe storm from the southwest moved over the Ohio valley on February 21, 1912, succeeded

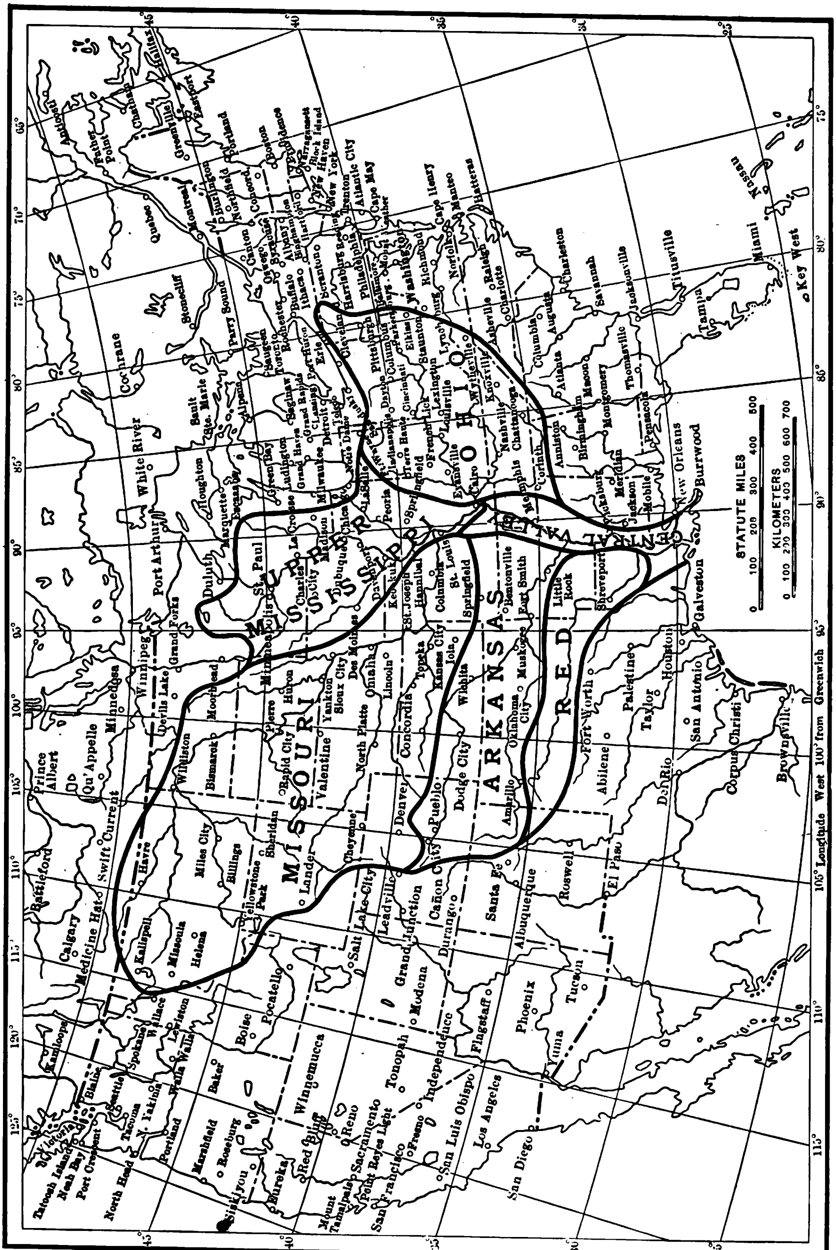


FIG. 95. DRAINAGE BASIN OF THE MISSISSIPPI RIVER

After U. S. Weather Bureau

by another storm of similar character four days later. During the first decade of March storms were frequent but not heavy. On March 10, an extensive depression moved in from the Pacific to southern California, and by the following day had reached Kansas. The rain and snow were moderately heavy. While this storm was moving across the country, another disturbance appeared on the North Pacific coast, and by the time (March 14) the first storm had passed to Newfoundland the second storm was over Kansas, accompanied, like its predecessor, by a secondary depression extending southward over southeastern Texas. This second storm passed into the North Atlantic Ocean during the night of March 15–16, and for four days there was no precipitation of consequence. On March 19 a depression was noted over Utah, following one over Kansas. The depression in the Great Basin was probably only a further development of the marked depression on the North Pacific coast of March 15–16, a fact which appears to have been overlooked by the writers on the floods of March, 1912. This storm passed into the North Atlantic during the night of March 21–22. On March 23 a well-marked storm appeared over the west Gulf, reached the Ohio valley on the 24th, and, passing rapidly to southern New England, moved thence to sea. It would seem then that there is a definite relation between storm frequency and floods.

The flood of 1912

The flood of March–April, 1913, which began on March 23, was caused solely by excessive precipitation over a large area. This heavy rainfall caused the rivers of northern Indiana and Ohio, especially the Miami, Scioto, and Muskingum, to rise rapidly. Only a small part of the great damage done can be attributed to the breaking of dams. These northern tributaries of the Ohio are not as a rule operative in causing floods in the Ohio. In this instance it happened, however, that the eastern and southern tributaries were also carrying large volumes of water. In an official report on this flood[1] it is stated that the almost inconceivably extensive damage done was increased by the work of man in the channels, along the banks, and

The flood of 1913

[1] Horton and Jackson, *Water Supply Paper No. 334*, U. S. Geological Survey.

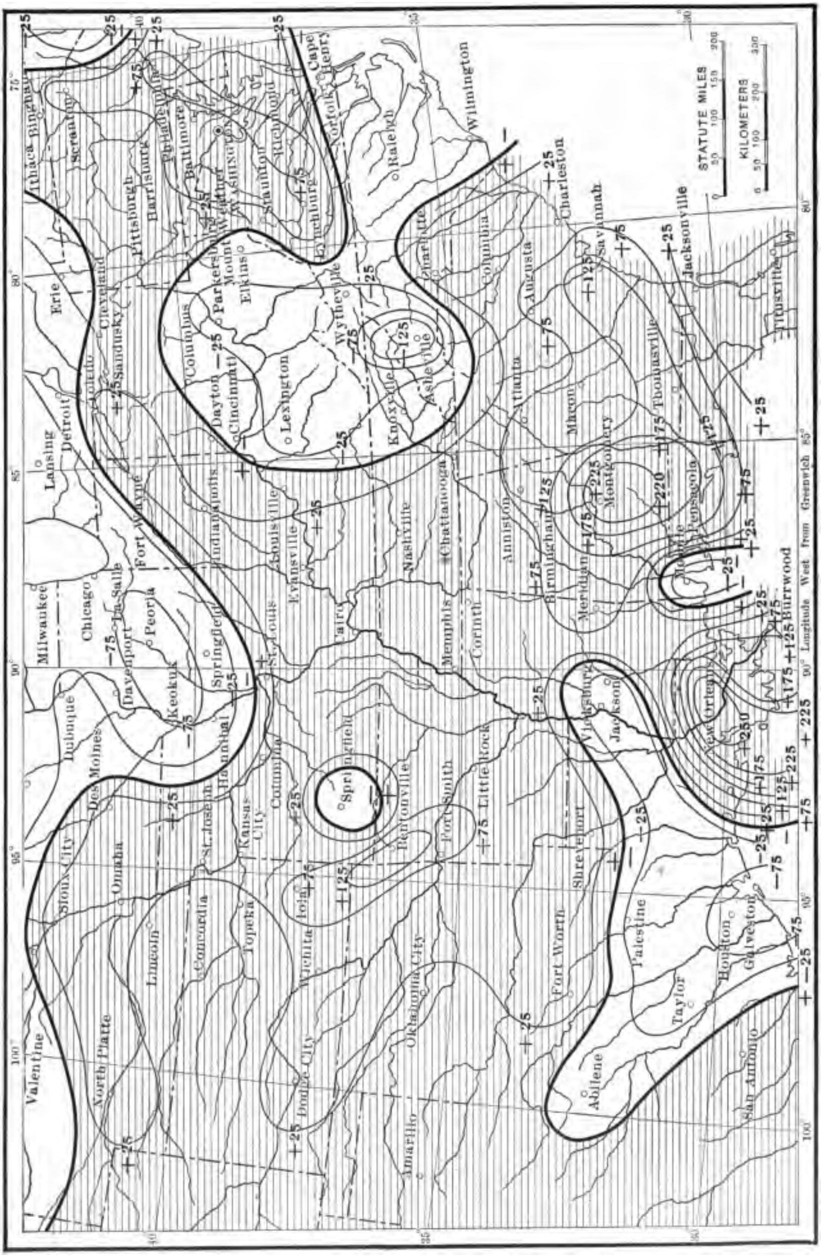


FIG. 96. PRECIPITATION DURING FLOOD

Departure from normal precipitation January 1 to April 2 inclusive, 1912, is given in millimeters.

From Water Supply Paper 334, U. S. Geol. Surv.

FIG. 97. MIAMI STREET CANAL BRIDGE, DAYTON, OHIO, AFTER THE FLOOD OF MARCH–APRIL, 1913

across the river valleys. The ground was not frozen; but it was practically saturated by previous rains, and so did not permit the storage of any considerable amount of water. It is doubtful, however, if ground storage even under the most favorable conditions would have had any material effect, because of the intensity of the precipitation. In the five days from March 23 to 27 the rainfall averaged from 100 to 250 millimeters. In advance of the first storm, which caused the tornadoes of the 23d, a marked hyperbar (area of permanent high pressure) drifted slowly across the United States, settling over the Bermudas on the 27th. Thus while a second storm was trying to move eastward during the 24th, there was an area of high pressure off the Atlantic coast, and another spreading eastward from the region of the Great Lakes. At 8 P.M. of the 24th these two areas were separated only by a lane of low pressure over the Ohio basin, connecting an approaching and a departing storm. Thus the rain areas of two storms came together and caused the most disastrous flood in the history of the Ohio valley.

The most disastrous flood of the Ohio valley

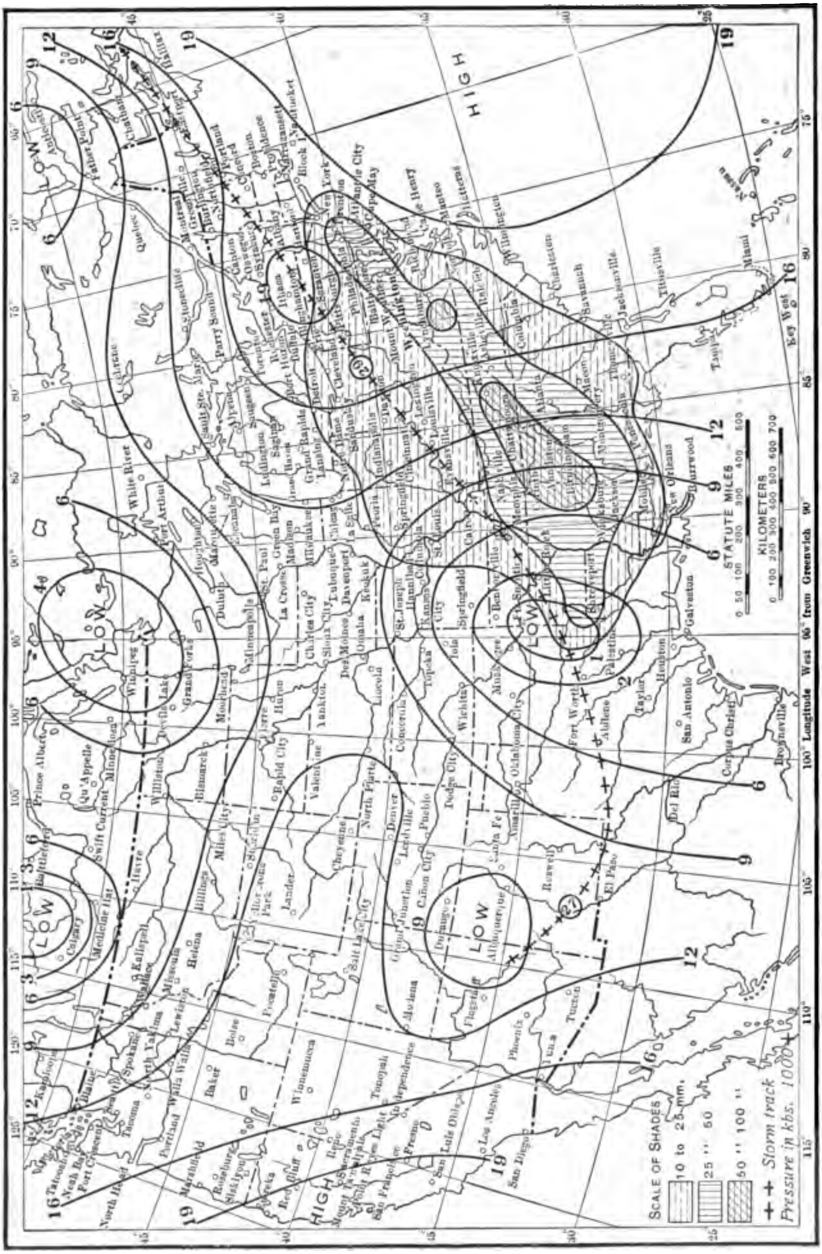


FIG. 98. WEATHER CONDITIONS, MARCH 28, 1912, AND PRECIPITATION DURING FOLLOWING 24 HOURS

From Water Supply Paper 334, U. S. Geol. Surv.

FIG. 99. POST-OFFICE, DAYTON, OHIO, AFTER THE FLOOD OF MARCH–APRIL, 1913

(Figs. 97 and 99.) No extremely low temperatures occurred during the flood; the ground was not frozen, and there was no snow or ice of any consequence stored in the drainage basin. The total damage, as estimated in the report of the engineers, was $180,000,000. The conditions at Dayton, Middletown, Hamilton, Piqua, Zanesville, and other localities were beyond description.

67. The Galveston storms. Perhaps the most destructive single storm of modern times occurred on the night of September 8, 1900, when a West Indian hurricane passed over Galveston. In that locality alone, more than 6,000 persons lost their lives through drowning or injury from falling buildings and storm damage. Property worth $30,000,000 was destroyed within the city limits and an enormous amount of property was ruined, in the interior and along the coast, with considerable loss of life. At Galveston a pressure reading as low as 964 kilobars (28.48 inches) was recorded, which was lower by 3 kilobars (.10 inch) than had been previously reported from any station in the United States. The highest wind

velocity was 38 meters per second (84 miles per hour) at 6:15 P.M., and 2 miles were registered at the rate of 45 meters per second (100 miles per hour). But at this time the anemometer was blown away and an estimate of the velocity of 50 m.p.s. (112 m.p.h.) seems not unreasonable. The devastation at Galveston was in large measure due to a wave which swept in from the Gulf in advance of the vortex of the storm. A detailed description of the hurricane is given by Cline in the *Monthly Weather Review* for September, 1900, p. 372. For an interval of nearly fifteen years, with the exception of a storm of moderate violence on July 21, 1909, this part of the coast of Texas was free from hurricanes. On August 16, 1915, a marked vortex approached the east Texas coast and by 8 P.M. of the next day the pressure at Galveston was down to 985 kilobars with high northeast winds. During the night

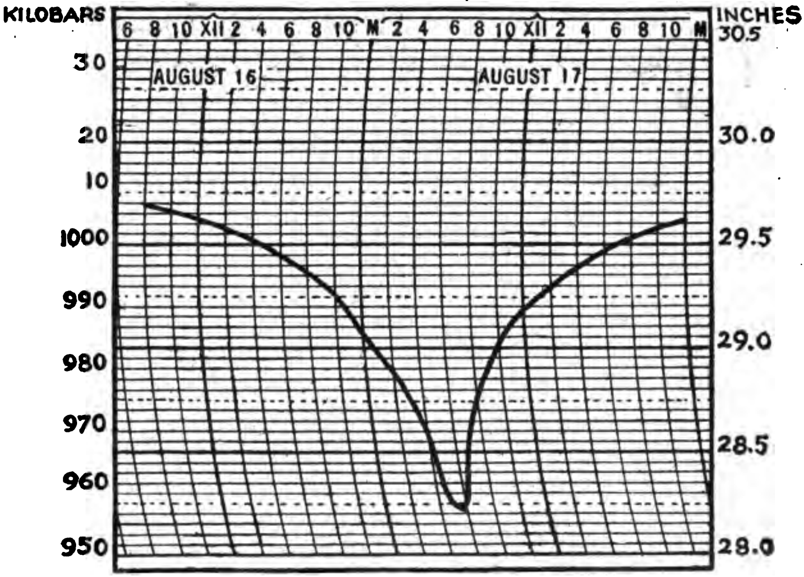
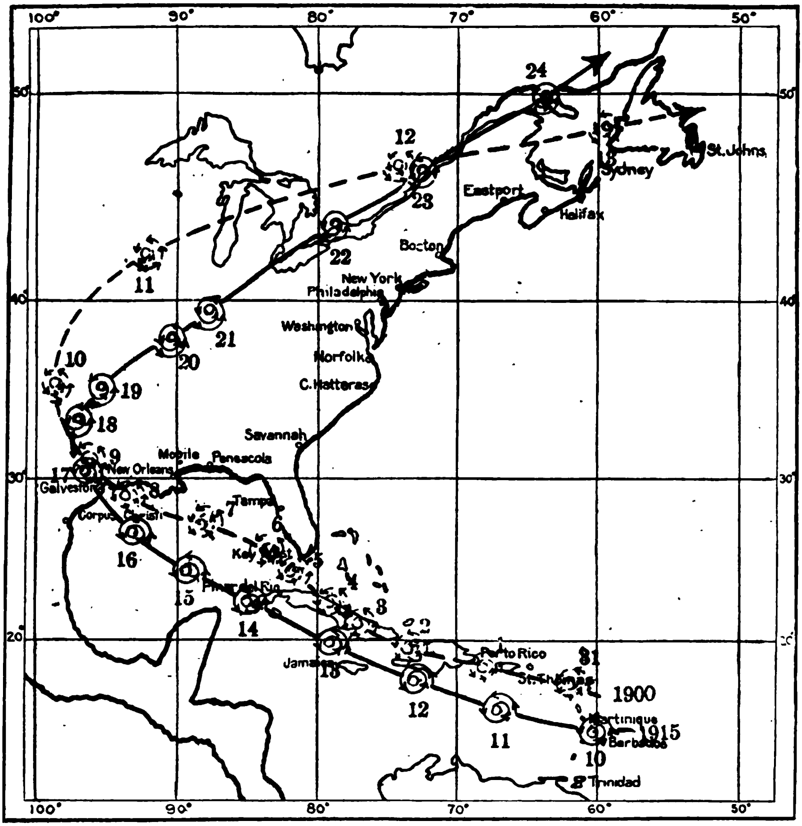


FIG. 100. BAROMETRIC PRESSURE, HOUSTON, TEXAS, DURING THE STORM OF AUGUST 16–17, 1915

of August 16–17 the storm passed slowly northwestward; and at 5:30 A.M. of the 17th the pressure at Houston (Fig. 100) was down to 955 kilobars[1] (28.20 inches), the lowest reading

[1] See p. 258, where a reading of 952 kbs. is given.



·FIG. 101. PATHS OF THE GALVESTON HURRICANES OF 1900 AND 1915

of the storm. Within a few hours the storm recurved to the northeast, moving slowly and accompanied by torrential rains and high winds. A comparison of the paths of the storms of 1900 and 1915 has been given by Frankenfield in a special bulletin from which the following extract is taken. (See Fig. 101, which shows the paths of the Galveston hurricanes of 1900 and 1915.)

"An inspection of these paths discloses the fact that the total time occupied from the first to the last appearance of both storms within the field of observation was exactly fourteen days, and that the storm of 1900 moved with a slower velocity of progression before reaching its recurve than after,

whereas in the storm of 1915 the reverse was true. The two paths are very similar in many respects, although that of 1915 lay a little to the southward of that of 1900 until the St. Lawrence Valley was reached. In previous published reports on the storm of 1900 the storm path shows a strong deflection toward the southwest Florida coast, but reports received from vessels and other sources after those publications indicated the fact that this deflection to the right was not so strong as has been supposed, and the track as here charted is thought to represent more nearly the true conditions. It was carefully plotted from all available observations. As to the comparative intensities of the two storms, it is perhaps idle to speculate. The wind velocities were not greatly different, and the effects of the two storms were much the same, except as modified by artificial

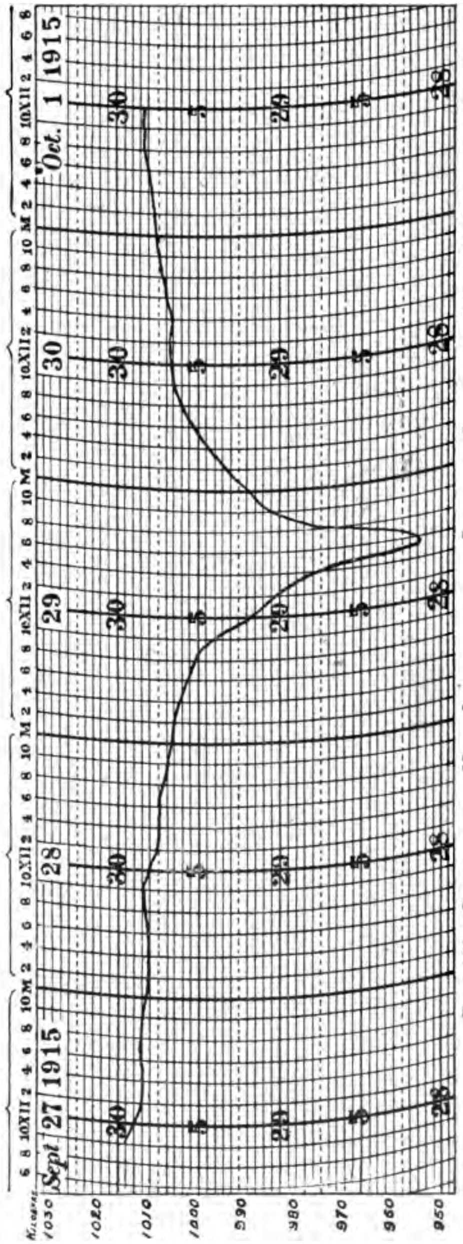


FIG. 102. PRESSURE AT NEW ORLEANS DURING STORM OF SEPTEMBER 29–30, 1915

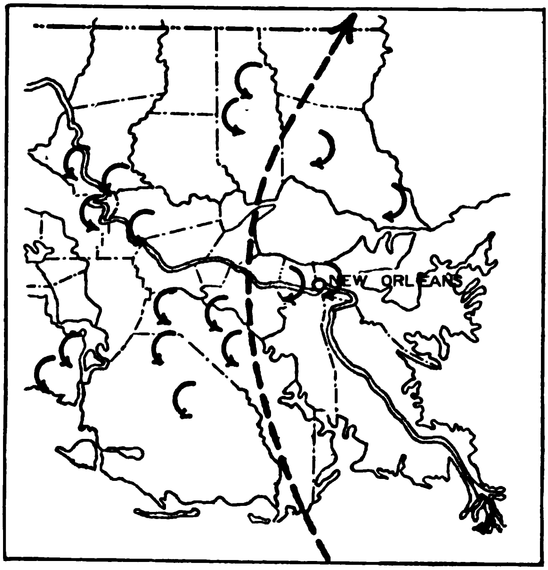


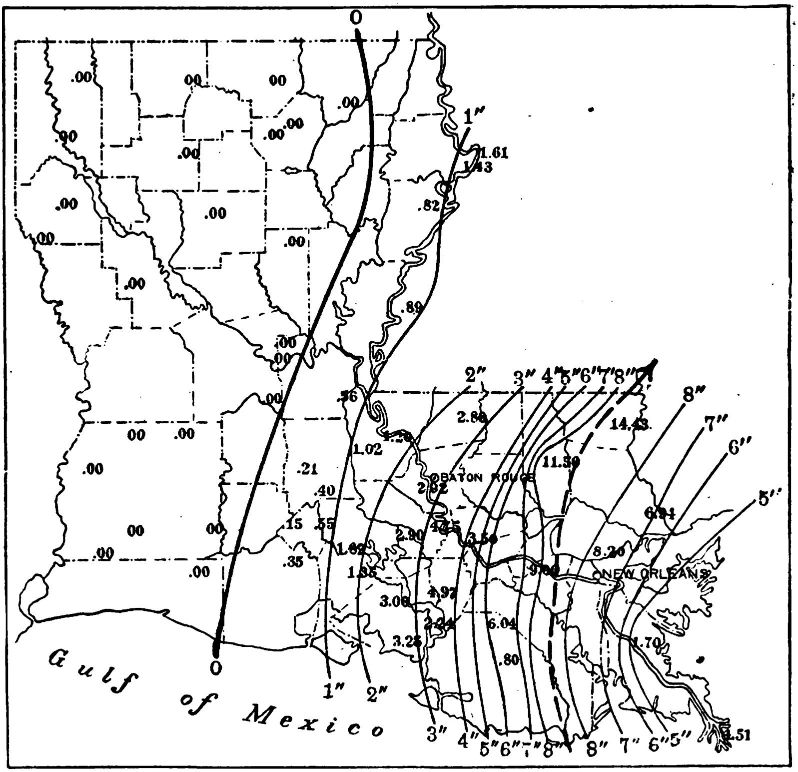
FIG. 103. CHANGE IN WINDS NEAR THE STORM CENTER

conditions in the vicinity of Galveston."

In the 1915 storm the loss of life was about 275, "whereas in 1900 the loss at Galveston and vicinity alone was at least 6,000. The great difference in favor of the storm of 1915 was due in greatest measure to the sea wall which was constructed by the city of Galveston shortly after the flood of 1900. There can be no question but that this wall saved the lives of thousands of people. It should also be remarked that the personal efforts of the official in charge of the local office of the Weather Bureau at Galveston were instrumental in saving the lives of hundreds of dwellers on Galveston Island. The official at Galveston sent out men on motorcycles to all places that could be reached on Galveston Island, who warned the inhabitants of the coming of the storm and impressed upon them the fact that unless they immediately, sought places of safety they would surely lose their lives. Subsequent occurrences confirmed the timeliness and correctness of this warning. Much commendation is also due to Prof. W. B. Stearns, coöperative observer and storm warning displayman at Seabrook, Tex. Upon the receipt of the first general warning on Sunday, August 15, and again Monday, August 16, Professor Stearns personally visited all the inhabitants at the low places in Seabrook and warned them to remove to places of safety on higher ground. There were formerly 88 houses on the bay front, and now there are 3.

Nevertheless, all of the former occupants were saved, except two, who apparently did not heed the warnings. Outside of Galveston the greatest loss in life probably occurred in the vicinity of Texas City, across the bay from Galveston."

68. The New Orleans storm of September 29, 1915. It is seldom that three hurricanes approach the Gulf Coast within a period of six weeks, yet such was the case in August to October in 1915. The Galveston storm of August described above was followed by a storm of moderate severity which passed



Cline

FIG. 104. RAINFALL DISTRIBUTION IN WEST INDIAN HURRICANE, SEPTEMBER 28, 29, 30, 1915

inland near the mouth of the Apalachicola River on September 4. This was followed by a storm of marked intensity which passed near New Orleans on September 29, described by

Cline in the *Monthly Weather Review* for September, 1915, as "the most intense hurricane of which we have record in the history of the Mexican Gulf Coast." An exhaustive and most instructive study of the storm has been made by Cline; and a more general discussion given by Bowie in a special Bulletin. The storm was first noted by the forecast officials on September 22 in the doldrums, in latitude 15° N. and longitude 64° W. By the night of September 29 the center was over New Orleans, surpassing in intensity the Galveston storm of August. The lowest pressure, reduced to sea level and corrected for standard gravity, was 952 kilobars (28.11 inches), which is the lowest reading ever recorded at a Weather Bureau station. The extreme wind velocity was approximately 58 meters per second (130 miles per hour) from the east. Fig. 102 shows the sea-level pressure at New Orleans during the passage of the storm; Fig. 103, the wind changes along the hurricane track; and Fig. 104, rainfall distribution in the various parishes of Louisiana near the path of the center.

Lowest pressure 952 kbs.

CHAPTER XVII

Frosts

69. The relation between the surface flow of air and frosts. Only in recent years have aërographers given much attention to the slow-moving currents of the lower strata of the atmosphere. These strata differ greatly from the whirls and cataracts of both high and low levels which we familiarly know as the "winds." The larger and more energetic air streams play a part in the formation of frost, and their importance in this regard is not to be underestimated. However, it is a slow surface flow, almost a creeping, of the air near the ground which chiefly controls the temperature there and is all-important in frost formation. It is, therefore, of some importance to study the conditions which bring about this slow movement or displacement of air. It is true that there are times when, owing to thorough mixing and ventilation, there is little opportunity for slow displacement; then the temperature will fall to low points and damage from frost result. But such conditions are more properly described as cold waves (though the term is somewhat misleading), or "freezes." In such cases there is an unusual loss of heat by direct convection and a transfer of masses of cold air. Strictly speaking, frosts are connected with temperature inversions brought about by a vertical, rather than a horizontal, movement of the air; and their problem is, therefore, essentially one of local air drainage. The expression "local air drainage" requires some defining. So far as known, it was first used by the author[1] in explaining frost. It was there shown that in the valleys of California a well-marked flow of the surface air can be traced and utilized in forecasting frosts. The condensed water vapor or fog can be seen drifting into the valleys or settling in the low places. There are well-marked stream lines, and one is led to believe

Chief factor in frost formation

Local air drainage

[1] "Frost Fighting," *Bulletin No. 29*, U. S. Weather Bureau, 1900.

that the mixture of air and water vapor of a given temperature, say 275°A., cools or is chilled by contact with the hilltops, and under the influence of gravity and other causes flows down the slopes. The term "air drainage" has been objected to. A recent writer has insisted that the flow of air down a hillside is not comparable to a flow of water, since water is an incompressible fluid where flow would be determined by gravity alone, while air is a compressible gas; and further because water in place may flow away, leaving the space it occupied vacant. But this writer forgets that, as von Bezold and others have shown, any substance will rise or fall without additional cooling or warming "*when it forms part of an endless chain* that glides frictionless over a roller and to which there has once been given a velocity, no matter how small."

The flow of air

It is well known that soon after sunset, valleys and low places serve as catchment basins for slow-moving air, denser and colder than that above. The hilltops, the terraces, and even the mountain tops, if not too high, are in contact with air of higher temperature, which must be either an indraft from warm surrounding strata or the displaced air from below. How the circulation begins and how it is maintained are not clearly understood; and, unfortunately, we have no instruments sufficiently sensitive to record this. The cooling of the lower levels, the warming of the upper levels, and the existence of an inversion are evidently not the result of a single cause. But one fact stands out strongly in all the investigations thus far made; and that is, where the air is in motion there is less likelihood of frost than where the air is stagnant.

Likelihood of frost greater where air is stagnant

70. The various processes in frost formation. One cause of circulation is the fact that the slopes, especially those facing west of southwest, have been heated by insolation during the day, and therefore radiate more rapidly. Radiation is a function of the absolute temperature, and, other things being equal, the surface that is warmest will radiate at a more rapid rate. The energy thus radiated is not absorbed by any layer of

Effects of radiation

vapor or by dust particles, as may be, and generally is, the case in the lower levels. The valleys and low levels also lose heat by radiation; but soon after sunset there is formed a thin blanket of condensed vapor, which interferes with free radiation and checks the rate of cooling. The air at the higher level is drained of its load of vapor and dust nuclei, becoming more and more like a pure gas and permitting freer radiation. A mixture of air and vapor per unit volume is lighter than dry air; and thus moist air naturally tends to rise. The condensed vapor, however, must be regarded in a different light from the vapor before condensation. In condensing, and also, but to a less

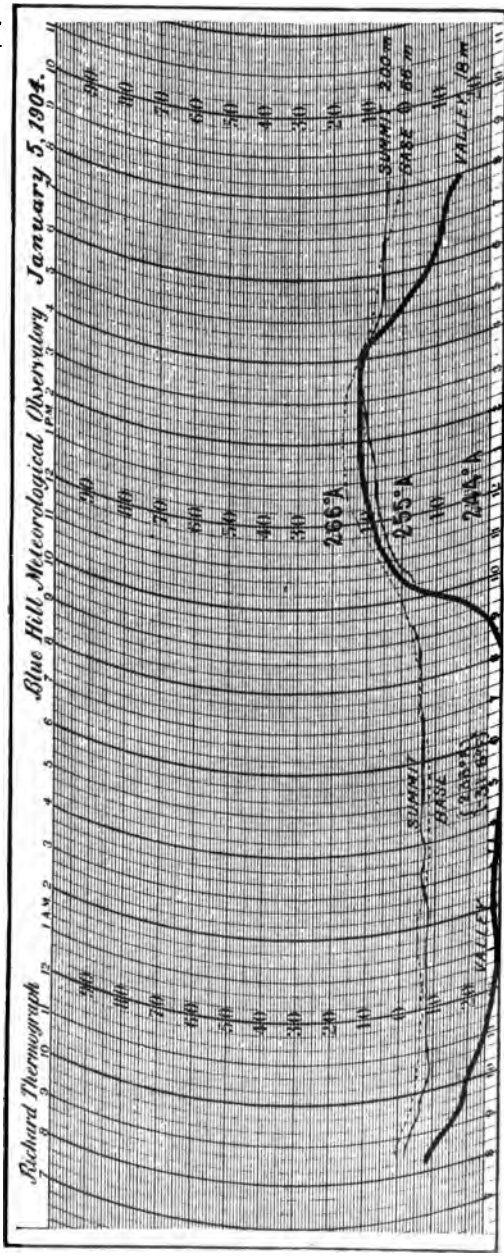


FIG. 105. TYPES OF INVERSION. WINTER TYPE
Marked cooling 8 P.M. to 12 midnight; marked warming 8 A.M. to 11 A.M.

degree, in congealing into frost flakes, heat is set free in the sense that molecular energy is decreased. This heat is not shown as a direct rise in temperature, but does serve to prevent any fall in temperature, such as expansion due to rising would produce. Thus we have near the ground an increasing load of condensed vapor, or vapor near the condensing point, which either crystallizes as frost with further cooling or is carried away by convective currents.

Formation and effect of condensed vapor

We see, then, that there are various conflicting processes: we have gain and loss of heat by radiation, the upper slopes losing heat by radiation and the lower air masses gaining heat; retardation or acceleration of rate of temperature change with the change in state of the water vapor; direct gain or loss of heat by convection, or the actual translation of cold and warm air masses; and, finally, some slight gain by conductivity.

Conflicting processes

Unfortunately, the word "frost" has been used as the equivalent for lowest temperature, whereas it is more properly simply an indication of the existence of sufficient water vapor, some of which has been changed into ice in the form of spicular crystals. Such deposit does not necessarily indicate the place of lowest temperature; for with other than saturation conditions, lower temperatures may prevail without the formation of crystals.

Water vapor in frost formation

Some good illustrations of the inversion of temperature are shown in the accompanying diagrams (Figs. 105, 106, and 107).

Fig. 105 illustrates a remarkable inversion which occurred January 5, 1904, when the temperature at the valley station at Blue Hill fell to 233°A. These records show how great the difference in temperature may be, at different levels on the side of a hill, during the early morning hours on a still winter day. This is the greatest inversion recorded at Blue Hill. There was also an inversion of similar character on the succeeding night. In Fig. 106 is shown an inversion occurring on February 25, 1914, which is of special interest, since the effective cause of cooling did not begin soon after sunset, as is the case with most

Temperature inversions

inversions. In fact, it did not manifest itself until long after midnight. The records illustrate how the air is stratified at times of frost, the coldest, densest air settling in the valley, and being displaced during the forenoon. Fig. 107 illustrates typical early fall and late spring inversions, of special interest to gardeners and truck farmers. In the records is shown the drainage of cold air to the lower levels and the likelihood of injurious frost while the higher levels are immune. This explains why frosts are so frequent in the lowlands as compared with the slopes. Figs. 108 and 109 show frost on glass.

In all these it will be noticed

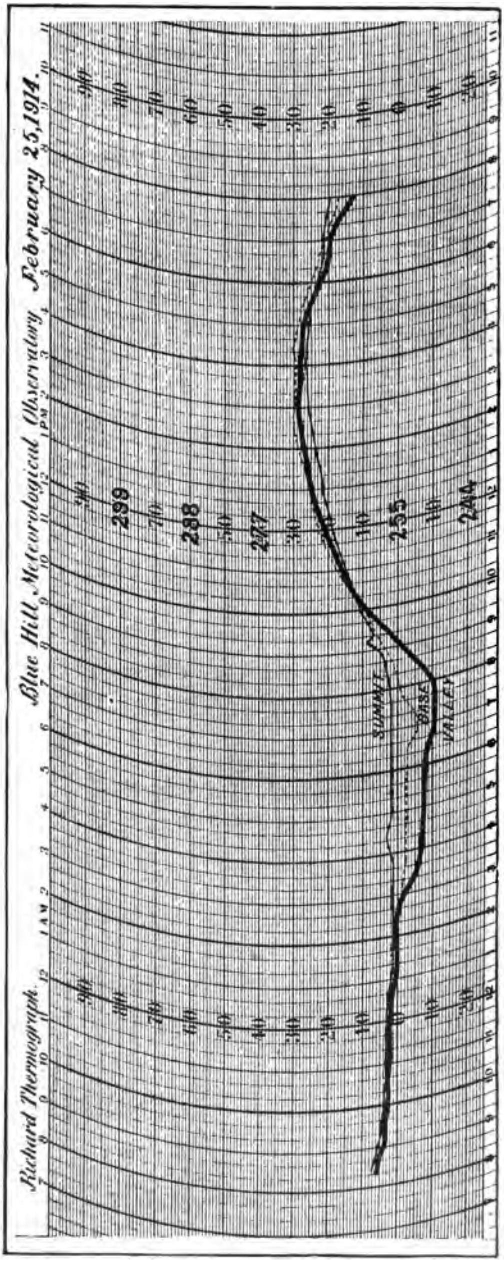


FIG. 106. TYPES OF INVERSION. UNUSUAL TYPE
Marked cooling began 2.20 A.M.

that there is a rapid rise in temperature at the lowest level, shortly after sunrise, and a slow rise at the base and a still slower rise at the summit. The respective heights of the three stations are: valley, 18 meters; base of hill, 66 meters; and summit, 200 meters above sea level. This rapid rise at the valley level is significant, for it indicates that with the formation of convective currents due to insolation there is air movement and effective heating. One might infer from this that the cause of cooling is not radiation, but the rapid dying out of convectional currents near the ground after sunset, any uprise of air being counterbalanced by the slow down-moving air from the higher slopes; not from the mass of air itself, which in the main is warmer, as we have seen, and which remains relatively warm throughout the night, since radiation from air is less rapid than from the hillsides. Apparently, then, there is a slow flow of air down the sides into the valley. It seems plain from these inversions that the principal reason why the summit temperature remains high during the night is because of the existence of a moderate air movement and consequent mixing. Frequent eye observation of the rate of ascent of smoke from a valley on quiet, clear afternoons may enable one to surmise that inversion already exists to some degree and thus accurately foretell night inversion and the frosts of the next morning. Any change in velocity or direction of air flow is accompanied with fluctuation in temperature.

Rise in temperature at the lowest level

High temperature at high levels due to mixing

There are easily recognizable certain types of storm movement which are followed by frosts. Gusty northwest winds, dying out at sunset, with unclouded skies and low and decreasing humidity above 100 meters but increasing in the lower levels, are significant local conditions favoring frost. Some writers have made use in their frost discussions of the adiabatic rate of fall of temperature, which is 0.98 degree for each hundred meters' rise; but no such conditions have been found to occur at times of frost. On the contrary, as we have seen above, there is gain and loss of heat in various

Types of storm movement preceding frosts

ways, and adiabatic equilibrium is out of the question. Neither the adiabatic rate for dry air nor that for saturated air holds from the ground up to 200 meters. Instead of a fall, there is a rise in temperature.

As the temperature rises the humidity falls, as a general rule. Unfortunately, our instruments for recording are unsatisfactory. Relative humidity, standing by itself and as ordinarily expressed, is very misleading; in fact, it means a ratio in which one term is suppressed. No proper study of frost or temperature inversion can be made without a full and definite knowledge of the behavior of the water vapor and dust content of

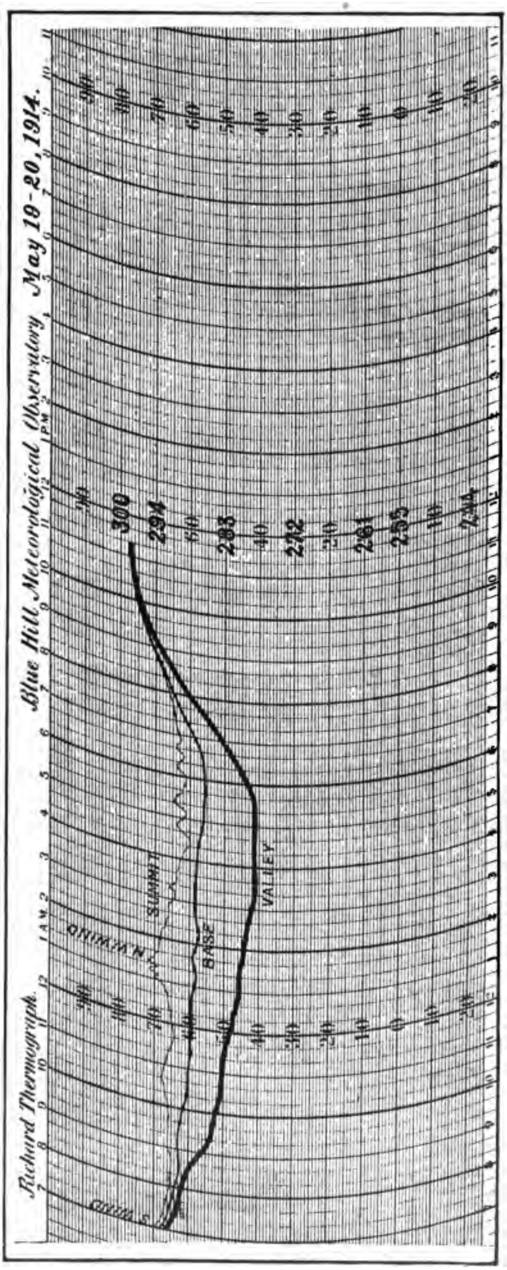


FIG. 107. FROSTS AND INVERSIONS. TYPICAL LATE SPRING FROST


McAdie

FIG. 108. FROST CRYSTALS

the atmosphere. A form of instrument devised by the author is a decided improvement over the usual form of hygrograph. It is known as a "saturation deficit recorder." It gives a continuous record of the weight of the vapor in grams per unit volume. The temperatures are given in degrees absolute, but the record sheet also is graduated to show the saturation weights for each degree. The thermograph part of the instrument (Fig. 110), therefore, records the appropriate weight of the water vapor per cubic meter at saturation, and the hygrograph part gives the existing percentage. The difference between the two is the saturation deficit, a quantity that may be used to advantage in discussions of frost formation or in the more general problem of the change in a given volume of moist air as it rises or falls or is transported

"Saturation deficit recorder"

from a region of high to a region of low pressure. It is regrettable that we have no method of recording continuously what von Bezold has termed the "mixing ratio" or the mass of vapor mixed with a unit mass of dry air expressed as a fraction of this latter unit; and that we have no way of ascertaining what the same investigator calls "specific humidity," or the quantity of vapor in a unit mass of moist air expressed in fractional parts of this unit. Moreover, there is a difference between vapor pressure and absolute humidity, although the two are often considered as equivalent. The record of the mixing ratio would be important in frost work, since, for any given change of level, the change in the mixing ratio would give the quantity of ice deposited, — not

Vapor pressure and absolute humidity

McAdie

Fig. 109. Frost Crystals

McAdie

FIG. 110. SATURATION DEFICIT RECORDER

This instrument records the temperature in degrees absolute, and the saturation weight of the vapor for given temperatures.

in this case, due to ascent of the air, but to contact with the hillsides cooled by rapid radiation and to the floor of the valley which also acts as a condensing surface. Moreover, the quantity of water (or frost crystals) thus separated from the air would give an indication of the intensity of the up-and-down movement of the air and its content of vapor. We have, indeed, in frost conditions a problem somewhat similar to that in certain cloud formations,— billows and bars,— less pronounced, but none the less phenomena of moving air strata of different temperatures and densities in close proximity. It may be pointed out, as Neuhoff has shown in his reconstruction of the Hertz diagram, that at 273°A., altitude lines run parallel to pressure lines at equal distances from each other for equal pressure changes; hence the isothermal change of altitude at freezing temperature is proportional to the quantity of water present. Practically 1 gram of freezing water corresponds to a change of level of 27 meters.

Close proximity of unlike strata

A more direct form of instrument is such as that shown in Fig. 111. There is continuously recorded not only the temperature but also the temperature of evaporation. The pressure of the aqueous vapor can be readily determined from the formula given by Ferrel[1] or obtained directly from *Smithsonian Physical Table No. 150*, 1914, p. 157.

[1] *Report of Chief Signal Officer*, 1886, App. 24.

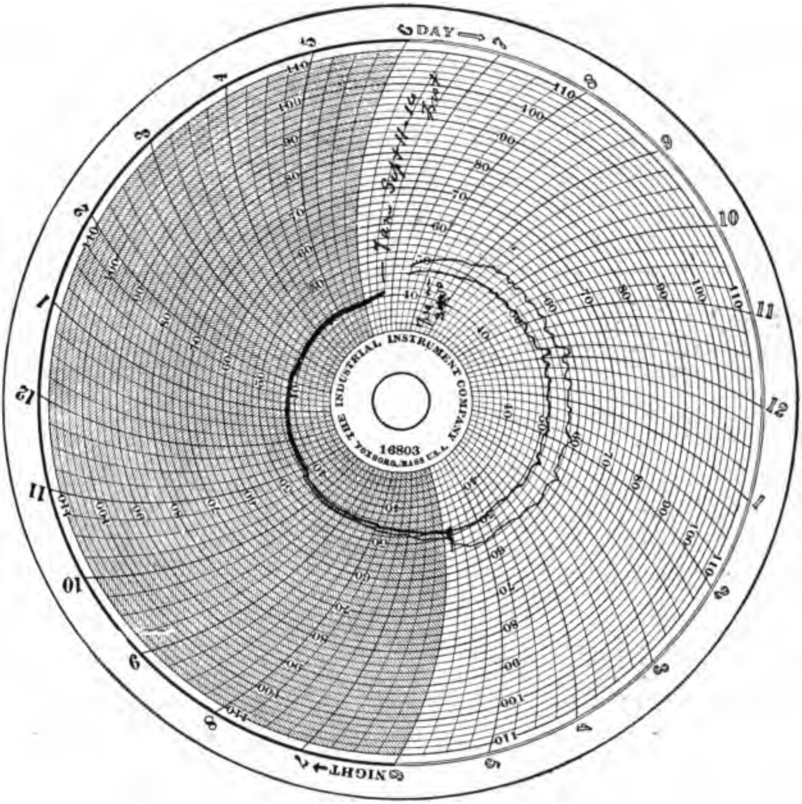


FIG. 111. THERMOGRAPH FOR DRY AND WET BULB READINGS
Foxboro type.

Dry and wet thermograph

The dew point, relative humidity, and humidity term 0.378, which occurs in the formula for density of air containing aqueous vapor at pressure, can be easily obtained. This instrument gives more accurate readings if the two thermometers are placed near a small fan or other ventilating device. It is preferable that this instrument be not inclosed in the usual louvered shelter. It should not be placed at the customary elevation of two meters above the ground, but as near the ground as possible.

It is also necessary to pay special attention to the purity of the water and the cleanness of the muslin used on the

wet bulb for evaporating the film of water. The pressure of saturated aqueous vapor varies somewhat at temperatures near 273°A., the variation depending upon whether the radiation is from a water or an ice surface. The following short table illustrates this difference:

Over water		Over ice	
Temperature °A.	Vapor pressure mm.	Temperature °A.	Vapor pressure mm.
270	3.67	270	3.56
271	3.95	271	3.88
272	4.25	272	4.21
273	4.58	273	4.58

We have seen that it is of importance to obtain reliable records of the amount of water vapor present, and, if possible, the changes which this quantity undergoes. One source of cooling is the abstraction of a considerable quantity of moisture from the air. A copious deposit of frost, however, does not necessarily indicate the region of lowest temperature.

71. Conversion table for frost work. Orchard heaters, evaporators, and frost protectors of various forms have come into such widespread use that a convenient table for the quick conversion of heat units into power units, and *vice versa*, seems to be much needed.

British thermal unit

It may be pointed out that the British thermal unit is the quantity of heat required to raise the temperature of 1 pound of pure water at maximum density, 39.1° F., 1° F. This is the unit frequently used by engineers in this country and Great Britain, *but it is desirable that the old English units and the Fahrenheit scale be used as little as possible.* A British thermal unit is equal to 0.252 calorie and also equal to 777.5 foot-pounds. One therm will raise the temperature of 1 gram of water 1° C.; 1,000 therms equal 1 calorie, equal to 3.968 British thermal units.

In problems connected with the heat of water, it should be remembered that the total heat is the latent heat plus the sensible heat. The total heat required to evaporate water

at a given temperature is 1,059.7 + 0.428 T, where T is a given temperature. This holds for temperatures between 273° A. and 373° A.

In changing to steam at 373°A. a pound of water at 373° A. absorbs 970.4 British thermal units and the total heat is therefore 1,150.4 British thermal units. This is starting from a temperature of 273° A. A pound of ice at 273° A. requires 142.4 British thermal units to change into water at 273° A.

The latent heat of aqueous vapor may be found from the following formula:

$$L_d = 1{,}091.7 - 0.572\ t_d$$

where

L_d = latent heat;
t_d = temperature of water.

For convenience in frost work the following may be used:

1 kilowatt hour = 3,412.66 B.T.U.
1 H.P. = 746.3 watts.
1 H.P. hour = 2,544.6 B.T.U.
1 B.T.U. = 777.5 foot-pounds.
1 B.T.U. = 0.252 calories.
1 calorie = 1,000 therms.
1 calorie = 3.968 B.T.U.
1 calorie per kilogram = 1.8 B.T.U. per pound.
1 pound of air at 32° F. occupies about 12.4 cubic feet.
1 pound of water at 212°F. occupies 0.0161 cubic feet.
1 pound of steam at 212° F. occupies 26.14 cubic feet.
1 pound of water at 212° F. contains 181.8 B.T.U.
1 pound of steam at 212° F. contains 1,150.4 B.T.U.
1 pound of ice requires 143.8 B.T.U. to change to water.
1 cubic foot of water at 212° F. weighs 59.84 pounds.
1 cubic foot of water at 62° F. weighs 62.2786 pounds.
1 cubic foot of steam at 212° F. weighs 0.03826 pound.
1 cubic foot of dry air at 32° F. weighs 568 grains.
1 cubic meter of dry air at 0° C. weighs 1,293.05 grams.
Specific heat of water, 1.
Specific heat of ice, 0.489.
Specific heat of water vapor, 0.453 at atmospheric temperatures.
Specific heat of air, 0.241.

Values given above are laboratory values, obtained by using distilled water. Ordinary drinking water is heavier

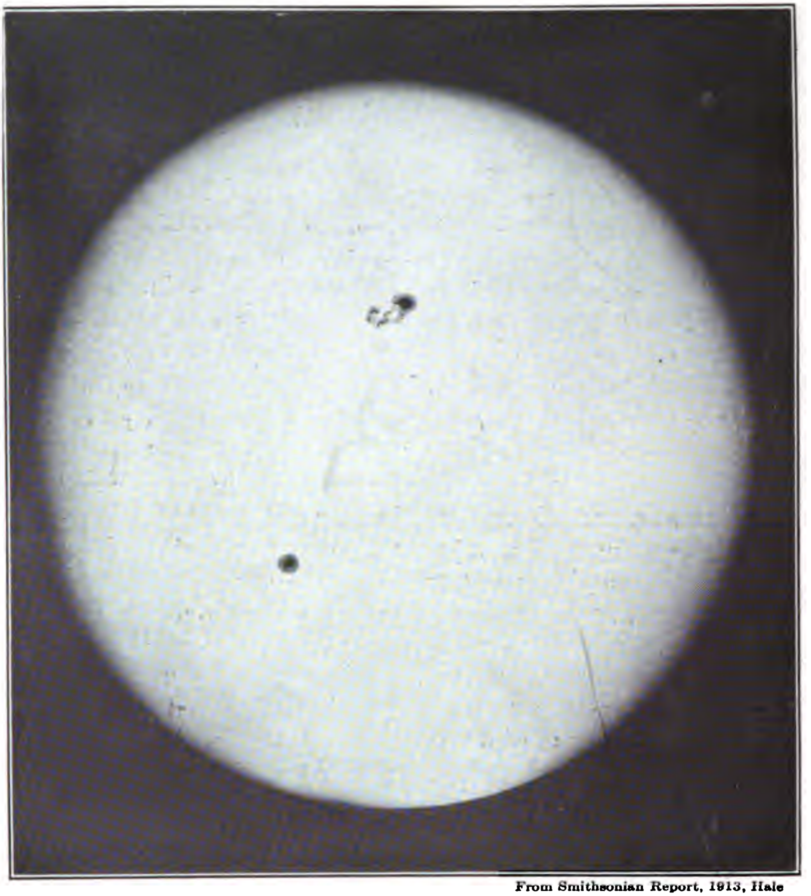
than distilled water, because of matter in solution. Salt water is also heavier. It may be remarked that the temperature of the freezing point in ordinary use, that is, 273° A., may not hold for the freezing of water in plant life. W. N. Shaw instances one plant where the freezing point is apparently 268°A. In other words, the change of water from the liquid to the solid state under natural conditions is somewhat different from the change as studied in a laboratory.

CHAPTER XVIII

Solar Influences

72. The source of radiant energy. The earth is, relatively speaking, an insignificant unit in the solar system. Fig. 112 illustrates the relative sizes of sun and earth. Furthermore, the solar system itself is an insignificant unit in the stellar universe. Astronomers tell us that the solar system is moving rapidly toward the constellation Hercules; but no appreciable effect upon the earth's atmosphere is known to result, nor is the amount of energy received from the stars sufficient to produce any observable effect. Efforts have been and are being made to measure the scattered radiation of the sky; but as yet there are no positive results. With the radiant energy of the sun, however, it is different; and here we have to deal with a prime mover. The mean distance of the sun from the earth is 149,500,000 kilometers or 92,900,000 miles. The solar parallax is 8.796 seconds and the sun's diameter 1,392,000 kilometers, or 865,000 miles. The velocity of light is 299,870 kilometers per second (186,300 miles), and the time required to traverse the mean radius of the earth's orbit is 498.8 seconds. The visible spectrum comprises light waves ranging in wave length from 0.7μ (0.0007 mm.), the red end, to $0.4\,\mu$ (0.0004 mm.), the violet end. At only a few aërological observatories are records of the intensity of solar radiation maintained. Perhaps one of the most serviceable records is that made at Davos, in the Swiss Alps, where continuous records have been obtained by C. Dorno. At this mountain station a continuous photographic record of the length of the ultra-violet spectrum (that is, the value of the shortest waves which penetrate the atmosphere) shows that the winter sun has great heating effect, but apparently does not attain a maximum in the other end of the spectrum. The spring sun has the greatest heat, with somewhat greater amount of ultra-violet radiation; the summer sun, much

Relative position of earth and sun



From Smithsonian Report, 1913, Hale

FIG. 112. DIRECT PHOTOGRAPH OF THE SUN

A dot one millimeter in diameter would represent the size of the earth.

heat and strongest utra-violet; and the autumn sun, much heat and much ultra-violet. Gockel thinks that herein lies the explanation of the "glacier burn," that is, an intense ultra-violet radiation. One point of interest is that the ultra-violet radiation undergoes more variation than the heat; and varies greatly with the season, so that a single day in summer may equal a winter month's total.

Heat and ultra-violet radiation

The intensity of solar radiation is measured by the heat

produced when a given surface exposed at right angles to the beam entirely absorbs the radiant energy. The mean value of the so-called "solar constant of radiation" has been fixed by Abbot at 1.932 calories per square centimeter per minute. This value differs materially from former values, especially the generally accepted value of 3 calories as given in many textbooks. If the solar constant were indeed constant, the earth would receive in a year something like one million million million million calories. In popular terms this would be sufficient heat to melt a layer of ice 33 meters thick over the entire surface of the earth annually, or to evaporate 1.66×10^{13} kilograms of water, provided there were no atmosphere, no absorption, and no reflection.

"Solar constant of radiation"

73. Variation in sunshine. At any given point there must, of course, be variation as the sun changes longitude. Thus on January 1, when the longitude is 1°, the ratio is 1.03; on March 1, longitude 59°, 1.02; on July 1, longitude 179°, 0.96; on September 1, longitude 240°, 0.98; and on December 1, longitude 330°, 1.03. Thus in winter the value is larger than in summer. The duration of sunshine can be determined for any given latitude from the hour angle converted into mean solar time and then multiplied by 2. Considering northerly declination positive, and southerly declination negative, we have for example in latitude 42° N. the following values:

Causes of variation in possible sunshine

Variation with longitude

DURATION OF SUNSHINE

DECLINATION OF THE SUN	LENGTH OF DAY	
	Hours	Minutes
−23° 27′	9	7
−20°	9	37
−15°	10	18
−10°	10	56
− 5°	11	33
0°	12	9
5°	12	45
10°	13	22
15°	14	1
20°	14	43
23° 27′	15	14

The greatest possible duration for other latitudes is:

Latitude	0	20°	40°	60°	66°	90°
Maximum insolation	$12^h\ 7'$	$13^h\ 20'$	$15^h\ 1'$	$18^h\ 52'$	24^h	6^{months}

If the unit of insolation be the amount received in a day at the equator on March 21, then for given latitudes values will vary in the following ratios:

Latitude	0	20°	40°	60°	North Pole	South Pole
March 21	1.0	0.93	0.76	.50	0	0
June 21	0.98	1.04	1.10	1.09	1.20	0
Sept. 23	0.88	0.94	0.70	0.30	0.	0
Dec. 21	0.94	0.68	0.35	0.	0.	1.28

Varying distance between earth and sun

The orbit which the sun appears to make around the earth, but which in reality is made by the earth around the sun, is not a circle but an ellipse inclined to the plane of the equator. The speed of the earth is not constant; and instead of traveling equal distances in equal times, the distance traveled is such as to make the areas swept over by the line joining earth and sun *equal in equal times.* So when the sun is nearest, the earth travels fastest. As we have said, the sun appears to travel in a plane which makes an angle of 23° with the plane of the equator.

Sunspot numbers

There may be other causes of variation in the intensity of solar radiation—changes which may be of solar origin and not periodic. Thus the monthly mean values of the solar constant from 1905 to 1912 have been compared with the so-called "Wolff sunspot numbers" for the same months, and it seems likely that increased values of the solar constant attend increased sunspot numbers. In the report of the Astrophysical Observatory for 1913 it is stated that there is an increase of radiation, at the earth's mean distance from the sun, of 0.07 calorie per square centimeter per minute with an increased spottedness of the sun, represented by 100 Wolff sunspot numbers.

Simultaneous observations at Mount Wilson and Bassour, Algeria, indicate that fluctuations in solar-constant values

found in California in earlier years may now be explained not as local phenomena but as due to causes outside of the earth; and thus we may conclude that the sun is a variable star, having not only a periodicity connected with the periodicity of sunspots, but also an irregular, non-periodic variation, sometimes running its course in a week or ten days, again in longer periods, and ranging over irregular fluctuations of from 2 to 10 per cent of the total. It has also been shown by Abbot, Fowle, Kimball, and others that great volcanic eruptions materially decrease the apparent solar radiation, or rather that atmospheric transmissibility undergoes marked changes with consequent diminution of temperature. Marked changes occurred in 1884–1886 (probably connected with Krakatau) and again in 1903–1904.

The sun a variable star

74. Measurement of solar radiation. By using a Callendar pyrheliometer and an eclipsing screen, the total radiation can be obtained in two components, one representing direct solar radiation and the other the diffuse sky radiation. The total radiation per square centimeter of horizontal surface with the clearest sky varies, according to Kimball, for the particular point of observation (near Washington), from 250 calories a day (December 20) to 765 calories (June 10). In general the radiation received on clear days during the first half of the year exceeds that of the second half by 8 per cent, probably due to the increased water-vapor content of the atmosphere during the latter period.

The total radiation received with the clearest sky *in mid-day* per square centimeter of horizontal surface varies from 45 calories in December to 90 in June. When clouds are near the sun but do not obscure it, the momentary maximum rates are increased by about 0.15 calorie.

The diffuse sky radiation received on a horizontal surface at noon averages about 25 per cent of that from the sun.

Expressed in units of work,

1 calorie per minute per cm.^{2} represents 697 watts per m^2.
90 calories per hour (1½ per minute) represent 1 kilowatt per m^2.

The radiation received on a square meter of horizontal surface

on a clear day in midsummer is, therefore, equivalent to 5 kilowatt hours.

Some recent measurements are:

American University, Washington, D.C., on December 24, 1914, with the sun at zenith distance 62.5°, an intensity of 1.48 calories per minute per square centimeter; and on February 28, 1915, with the sun at zenith distance 57.5° the intensity was 1.50 calories. At Santa Fé, N.M., elevation 2,133 meters, and in an arid region, a maximum of 1.64 calories was recorded with a zenith distance of 55°. In brief, at sea level, in summer and at mid-day, there reaches the earth each second .0225 calorie per square centimeter; and of this .0096 calorie is scattered or absorbed and .006 calorie re-radiated from the atmosphere. The amount of energy varies inversely as the square of the distance from the sun also with the angle of incidence of the rays, and according to duration.

Other radiations

It is possible that there is in the upper atmosphere a layer of cosmical dust which is strongly radioactive. Simpson has recently pointed out that the measurements of Vegard and Störmer on the aurora indicate true radioactive radiation penetrating the atmosphere and producing the same results as if the atmosphere were being bombarded from the outside by the α radiation, which is now being studied in so many physical laboratories. Experiments on ionization made in balloons in 1914 show the existence of a strong radiation. This may help explain the nature of the aurora. The average height of the bottom edge of the aurora as determined by 1920 measurements in Norway is 108 kilometers, and no aurora lower than 85 kilometers was noticed. It would seem that the cosmic rays producing the aurora are in two groups with different penetrating power. The diffuse arcs, the drapery, and more intense displays seem to be of the same physical nature.

Pyranometer

A pyranometer[1] is an instrument adapted to measuring heat coming from or going to space above. It was devised by Abbot, Aldrich, and Kramer of the Astrophysical Observatory of the Smithsonian Institution, for measuring accurately the intensity of sky light by day and of radiation outward toward the whole sky by night.

[1] For complete description see Smithsonian Misc. Coll., Vol. 66, No. 7, May, 1916.

APPENDIX

TABLE 1. INCHES INTO MILLIMETERS

1 inch = 25.40005 mm.

Inches	.00	.01	.02	.03	.04	.05	.06	.07	.08	.09
	mm.	mm.	mm.	mm.	mm.	mm.	mm.	mm.	mm.	mm.
0.00	0.00	0.25	0.51	0.76	1.02	1.27	1.52	1.78	2.03	2.29
0.10	2.54	2.79	3.05	3.30	3.56	3.81	4.06	4.32	4.57	4.83
0.20	5.08	5.33	5.59	5.84	6.10	6.35	6.60	6.86	7.11	7.37
0.30	7.62	7.87	8.13	8.38	8.64	8.89	9.14	9.40	9.65	9.91
0.40	10.16	10.41	10.67	10.92	11.18	11.43	11.68	11.94	12.19	12.45
0.50	12.70	12.95	13.21	13.46	13.72	13.97	14.22	14.48	14.73	14.99
0.60	15.24	15.49	15.75	16.00	16.26	16.51	16.76	17.02	17.27	17.53
0.70	17.78	18.03	18.29	18.54	18.80	19.05	19.30	19.56	19.81	20.07
0.80	20.32	20.57	20.83	21.08	21.34	21.59	21.84	22.10	22.35	22.61
0.90	22.86	23.11	23.37	23.62	23.88	24.13	24.38	24.64	24.89	25.15
1.00	25.40									

TABLE 2. FEET INTO METERS

1 foot = 0.3048006 meter

Feet	0	1	2	3	4	5	6	7	8	9
	m.	m.	m.	m.	m.	m.	m.	m.	m.	m.
0	0.000	0.305	0.610	0.914	1.219	1.524	1.829	2.134	2.438	2.743
10	3.048	3.353	3.658	3.962	4.267	4.572	4.877	5.182	5.486	5.791
20	6.096	6.401	6.706	7.010	7.315	7.620	7.925	8.230	8.534	8.839
30	9.144	9.449	9.754	10.058	10.363	10.668	10.973	11.278	11.582	11.887
40	12.192	12.497	12.802	13.106	13.411	13.716	14.021	14.326	14.630	14.935
50	15.240	15.545	15.850	16.154	16.459	16.764	17.069	17.374	17.678	17.983
60	18.288	18.593	18.898	19.202	19.507	19.812	20.117	20.422	20.726	21.031
70	21.336	21.641	21.946	22.250	22.555	22.860	23.165	23.470	23.774	24.079
80	24.384	24.689	24.994	25.298	25.603	25.908	26.213	26.518	26.822	27.127
90	27.432	27.737	28.042	28.346	28.651	28.956	29.261	29.566	29.870	30.175
	0	**10**	**20**	**30**	**40**	**50**	**60**	**70**	**80**	**90**
100	30.48	33.53	36.58	39.62	42.67	45.72	48.77	51.82	54.86	57.91
200	60.96	64.01	67.06	70.10	73.15	76.20	79.25	82.30	85.34	88.39
300	91.44	94.49	97.54	100.58	103.63	106.68	109.73	112.78	115.82	118.87
400	121.92	124.97	128.02	131.06	134.11	137.16	140.21	143.26	146.30	149.35
500	152.40	155.45	158.50	161.54	164.59	167.64	170.69	173.74	176.78	179.83
600	182.88	185.93	188.98	192.02	195.07	198.12	201.17	204.22	207.26	210.31
700	213.36	216.41	219.46	222.50	225.55	228.60	231.65	234.70	237.74	240.79
800	243.84	246.89	249.94	252.98	256.03	259.08	262.13	265.18	268.22	271.27
900	274.32	277.37	280.42	283.46	286.51	289.56	292.61	295.66	298.70	301.75
1000	304.80									

Table 3. Miles into Kilometers

1 mile = 1.609347 kilometers

Miles	0	1	2	3	4	5	6	7	8	9
	km.	km.	km.	km.	km.	km.	km.	km.	km.	km.
0	0	2	3	5	6	8	10	11	13	14
10	16	18	19	21	23	24	26	27	29	31
20	32	34	35	37	39	40	42	43	45	47
30	48	50	51	53	55	56	58	60	61	63
40	64	66	68	69	71	72	74	76	77	79
50	80	82	84	85	87	89	90	92	93	95
60	97	98	100	101	103	105	106	108	109	111
70	113	114	116	117	119	121	122	124	126	127
80	129	130	132	134	135	137	138	140	142	143
90	145	146	148	150	151	153	154	156	158	159
100	161									

Table 4. Kilometers into Miles

1 kilometer = 0.621370 mile

Kilometers	0	1	2	3	4	5	6	7	8	9
	Miles	Miles	Miles	Miles	Miles	Miles	Miles	Miles	Miles	Miles
0	0.0	0.6	1.2	1.9	2.5	3.1	3.7	4.3	5.0	5.6
10	6.2	6.8	7.5	8.1	8.7	9.3	9.9	10.6	11.2	11.8
20	12.4	13.0	13.7	14.3	14.9	15.5	16.2	16.8	17.4	18.0
30	18.6	19.3	19.9	20.5	21.1	21.7	22.4	23.0	23.6	24.2
40	24.9	25.5	26.1	26.7	27.3	28.0	28.6	29.2	29.8	30.4
50	31.1	31.7	32.3	32.9	33.6	34.2	34.8	35.4	36.0	36.7
60	37.3	37.9	38.5	39.1	39.8	40.4	41.0	41.6	42.3	42.9
70	43.5	44.1	44.7	45.4	46.0	46.6	47.2	47.8	48.5	49.1
80	49.7	50.3	51.0	51.6	52.2	52.8	53.4	54.1	54.7	55.3
90	55.9	56.5	57.2	57.8	58.4	59.0	59.7	60.3	60.9	61.5
100	62.1									

TABLE 5. INTERCONVERSION OF NAUTICAL AND STATUTE MILES

1 nautical mile* = 6080.27 feet

Nautical Miles	Statute Miles	Statute Miles	Nautical Miles
1	1.1516	1	0.8684
2	2.3031	2	1.7368
3	3.4547	3	2.6052
4	4.6062	4	3.4736
5	5.7578	5	4.3420
6	6.9093	6	5.2104
7	8.0609	7	6.0788
8	9.2124	8	6.9472
9	10.3640	9	7.8155

* As defined by the United States Coast Survey.

TABLE 6. CONVERSION OF VELOCITIES

Miles per hour into meters per second, feet per second, and kilometers per hour

Miles per hour	Meters per second	Feet per second	Kilometers per hour	Miles per hour	Meters per second	Feet per second	Kilometers per hour	Miles per hour	Meters per second	Feet per second	Kilometers per hour
0.0	0.0	0.0	0.0	**12.0**	5.4	17.6	19.3	**24.0**	10.7	35.2	38.6
0.5	0.2	0.7	0.8	12.5	5.6	18.3	20.1	24.5	11.0	35.9	39.4
1.0	0.4	1.5	1.6	13.0	5.8	19.1	20.9	25.0	11.2	36.7	40.2
1.5	0.7	2.2	2.4	13.5	6.0	19.8	21.7	25.5	11.4	37.4	41.0
2.0	0.9	2.9	3.2	14.0	6.3	20.5	22.5	26.0	11.6	38.1	41.8
2.5	1.1	3.7	4.0	14.5	6.5	21.3	23.3				
3.0	1.3	4.4	4.8	**15.0**	6.7	22.0	24.1	**26.0**	11.6	38.1	41.8
3.5	1.6	5.1	5.6	15.5	6.9	22.7	24.9	26.5	11.8	38.9	42.6
4.0	1.8	5.9	6.4	16.0	7.2	23.5	25.7	27.0	12.1	39.6	43.5
4.5	2.0	6.6	7.2	16.5	7.4	24.2	26.6	27.5	12.3	40.3	44.3
5.0	2.2	7.3	8.0	17.0	7.6	24.9	27.4	28.0	12.5	41.1	45.1
5.5	2.5	8.1	8.9	17.5	7.8	25.7	28.2	28.5	12.7	41.8	45.9
6.0	2.7	8.8	9.7	**18.0**	8.0	26.4	29.0	**29.0**	13.0	42.5	46.7
6.5	2.9	9.5	10.5	18.5	8.3	27.1	29.8	29.5	13.2	43.3	47.5
7.0	3.1	10.3	11.3	19.0	8.5	27.9	30.6	30.0	13.4	44.0	48.3
7.5	3.4	11.0	12.1	19.5	8.7	28.6	31.4	30.5	13.6	44.7	49.1
8.0	3.6	11.7	12.9	20.0	8.9	29.3	32.2	31.0	13.9	45.5	49.9
8.5	3.8	12.5	13.7	20.5	9.2	30.1	33.0	31.5	14.1	46.2	50.7
9.0	4.0	13.2	14.5	**21.0**	9.4	30.8	33.8	**32.0**	14.3	46.9	51.5
9.5	4.2	13.9	15.3	21.5	9.6	31.5	34.6	32.5	14.5	47.7	52.3
10.0	4.5	14.7	16.1	22.0	9.8	32.3	35.4	33.0	14.8	48.4	53.1
10.5	4.7	15.4	16.9	22.5	10.1	33.0	36.2	33.5	15.0	49.1	53.9
11.0	4.9	16.1	17.7	23.0	10.3	33.7	37.0	34.0	15.2	49.9	54.7
11.5	5.1	16.9	18.5	23.5	10.5	34.5	37.8	34.5	15.4	50.6	55.5

TABLE 6. CONVERSION OF VELOCITIES—*Continued*

Miles per hour into meters per second, feet per second, and kilometers per hour

Miles per hour	Meters per second	Feet per second	Kilometers per hour	Miles per hour	Meters per second	Feet per second	Kilometers per hour	Miles per hour	Meters per second	Feet per second	Kilometers per hour
35.0	15.6	51.3	56.3	**50.0**	22.4	73.3	80.5	**64.0**	28.6	93.9	103.0
35.5	15.9	52.1	57.1	50.5	22.6	74.1	81.3	64.5	28.8	94.6	103.8
36.0	16.1	52.8	57.9	51.0	22.8	74.8	82.1	65.0	29.1	95.3	104.6
36.5	16.3	53.5	58.7	51.5	23.0	75.5	82.9	65.5	29.3	96.1	105.4
37.0	16.5	54.3	59.5	52.0	23.2	76.3	83.7	66.0	29.5	96.8	106.2
37.5	16.8	55.0	60.4					66.5	29.7	97.5	107.0
38.0	17.0	55.7	61.2	**52.0**	23.2	76.3	83.7	**67.0**	30.0	98.3	107.8
38.5	17.2	56.5	62.0	52.5	23.5	77.0	84.5	67.5	30.2	99.0	108.6
39.0	17.4	57.2	62.8	53.0	23.7	77.7	85.3	68.0	30.4	99.7	109.4
39.5	17.7	57.9	63.6	53.5	23.9	78.5	86.1	68.5	30.6	100.5	110.2
40.0	17.9	58.7	64.4	54.0	24.1	79.2	86.9	69.0	30.8	101.2	111.0
40.5	18.1	59.4	65.2	54.5	24.4	79.9	87.7	69.5	31.1	101.9	111.8
41.0	18.3	60.1	66.0	**55.0**	24.6	80.7	88.5	**70.0**	31.3	102.7	112.7
41.5	18.6	60.9	66.8	55.5	24.8	81.4	89.3	70.5	31.5	103.4	113.5
42.0	18.8	61.6	67.6	56.0	25.0	82.1	90.1	71.0	31.7	104.1	114.3
42.5	19.0	62.3	68.4	56.5	25.3	82.9	90.9	71.5	32.0	104.9	115.1
43.0	19.2	63.1	69.2	57.0	25.5	83.6	91.7	72.0	32.2	105.6	115.9
43.5	19.4	63.8	70.0	57.5	25.7	84.3	92.5	72.5	32.4	106.3	116.7
44.0	19.7	64.5	70.8	**58.0**	25.9	85.1	93.3	**73.0**	32.6	107.1	117.5
44.5	19.9	65.3	71.6	58.5	26.2	85.8	94.1	73.5	32.9	107.8	118.3
45.0	20.1	66.0	72.4	59.0	26.4	86.5	95.0	74.0	33.1	108.5	119.1
45.5	20.3	66.7	73.2	59.5	26.6	87.3	95.8	74.5	33.3	109.3	119.9
46.0	20.6	67.5	74.0	60.0	26.8	88.0	96.6	75.0	33.5	110.0	120.7
46.5	20.8	68.2	74.8	60.5	27.0	88.7	97.4	75.5	33.8	110.7	121.5
47.0	21.0	68.9	75.6	**61.0**	27.3	89.5	98.2	**76.0**	34.0	111.5	122.3
47.5	21.2	69.7	76.4	61.5	27.5	90.2	99.0	76.5	34.2	112.2	123.1
48.0	21.5	70.4	77.2	62.0	27.7	90.9	99.8	77.0	34.4	112.9	123.9
48.5	21.7	71.1	78.1	62.5	27.9	91.7	100.6	77.5	34.6	113.7	124.7
49.0	21.9	71.9	78.9	63.0	28.2	92.4	101.4	78.0	34.9	114.4	125.5
49.5	22.1	72.6	79.7	63.5	28.4	93.1	102.2				

TABLE 7. PRESSURE

Inches of Mercury at 273°A. and 45° latitude, to Kilobars

For brevity, the fundamental equations may be written:

$g_{45} = 980.624$ cm/sec^2.

density of mercury at normal freezing-point of water = 13.5959.

1 mercury-inch = 33.8660 kilobars; 1 millimeter = 1.33320 kilobars.

1000 kilobars = 29.5306 mercury-inches = 750.076 millimeters.

Inches and Tenths	.00	.01	.02	.03	.04	.05	.06	.07	.08	.09
	KILOBARS									
27.0	914.3	914.6	915.0	915.3	915.7	916.0	916.3	916.7	917.0	917.4
27.1	917.7	918.0	918.4	918.7	919.0	919.4	919.7	920.1	920.4	920.7
27.2	921.1	921.4	921.8	922.1	922.4	922.8	923.1	923.4	923.8	924.1
27.3	924.5	924.8	925.1	925.5	925.8	926.2	926.5	926.8	927.2	927.5
27.4	927.9	928.2	928.5	928.9	929.2	929.5	929.9	930.2	930.6	930.9
27.5	931.2	931.6	931.9	932.3	932.6	932.9	933.3	933.6	933.9	934.3
27.6	934.6	935.0	935.3	935.6	936.0	936.3	936.7	937.0	937.3	937.7
27.7	938.0	938.3	938.7	939.0	939.4	939.7	940.0	940.4	940.7	941.1
27.8	941.4	941.7	942.1	942.4	942.8	943.1	943.4	943.8	944.1	944.4
27.9	944.8	945.1	945.5	945.8	946.1	946.5	946.8	947.2	947.5	947.8
28.0	948.2	948.5	948.8	949.2	949.5	949.9	950.2	950.5	950.9	951.2
28.1	951.6	951.9	952.2	952.6	952.9	953.2	953.6	953.9	954.3	954.6
28.2	954.9	955.3	955.6	956.0	956.3	956.6	957.0	957.3	957.7	958.0
28.3	958.3	958.7	959.0	959.3	959.7	960.0	960.4	960.7	961.0	961.4
28.4	961.7	962.1	962.4	962.7	963.1	963.4	963.7	964.1	964.4	964.8
28.5	965.1	965.4	965.8	966.1	966.5	966.8	967.1	967.5	967.8	968.1
28.6	968.5	968.8	969.2	969.5	969.8	970.2	970.5	970.9	971.2	971.5
28.7	971.9	972.2	972.6	972.9	973.2	973.6	973.9	974.2	974.6	974.9
28.8	975.3	975.6	975.9	976.3	976.6	977.0	977.3	977.6	978.0	978.3
28.9	978.6	979.0	979.3	979.7	980.0	980.3	980.7	981.0	981.4	981.7
29.0	982.0	982.4	982.7	983.0	983.4	983.7	984.1	984.4	984.7	985.1
29.1	985.4	985.8	986.1	986.4	986.8	987.1	987.5	987.8	988.1	988.5
29.2	988.8	989.1	989.5	989.8	990.2	990.5	990.8	991.2	991.5	991.9
29.3	992.2	992.5	992.9	993.2	993.5	993.9	994.2	994.6	994.9	995.2
29.4	995.6	995.9	996.3	996.6	996.9	997.3	997.6	997.9	998.3	998.6
29.5	999.0	999.3	999.6	1000.0	1000.3	1000.7	1001.0	1001.3	1001.7	1002.0
29.6	1002.4	1002.7	1003.0	1003.4	1003.7	1004.0	1004.4	1004.7	1005.1	1005.4
29.7	1005.7	1006.1	1006.4	1006.8	1007.1	1007.4	1007.8	1008.1	1008.4	1008.8
29.8	1009.1	1009.5	1009.8	1010.1	1010.5	1010.8	1011.2	1011.5	1011.8	1012.2
29.9	1012.5	1012.8	1013.2	1013.5	1013.9	1014.2	1014.5	1014.9	1015.2	1015.6
30.0	1015.9	1016.2	1016.6	1016.9	1017.3	1017.6	1017.9	1018.3	1018.6	1018.9
30.1	1019.3	1019.6	1020.0	1020.3	1020.6	1021.0	1021.3	1021.7	1022.0	1022.3
30.2	1022.7	1023.0	1023.3	1023.7	1024.0	1024.4	1024.7	1025.0	1025.4	1025.7
30.3	1026.1	1026.4	1026.7	1027.1	1027.4	1027.7	1028.1	1028.4	1028.8	1029.1
30.4	1029.4	1029.8	1030.1	1030.5	1030.8	1031.1	1031.5	1031.8	1032.2	1032.5

Table 7. Pressure—*Continued*

Inches of Mercury at 273°A. and 45° latitude, to Kilobars

Inches and Tenths	.00	.01	.02	.03	.04	.05	.06	.07	.08	.09
	KILOBARS									
30.5	1032.8	1033.2	1033.5	1033.8	1034.2	1034.5	1034.9	1035.2	1035.5	1035.9
30.6	1036.2	1036.6	1036.9	1037.2	1037.6	1037.9	1038.2	1038.6	1038.9	1039.3
30.7	1039.6	1039.9	1040.3	1040.6	1041.0	1041.3	1041.6	1042.0	1042.3	1042.6
30.8	1043.0	1043.3	1043.7	1044.0	1044.3	1044.7	1045.0	1045.4	1045.7	1046.0
30.9	1046.4	1046.7	1047.1	1047.4	1047.7	1048.1	1048.4	1048.7	1049.1	1049.4

Thousandths of an Inch

Inch	.001	.002	.003	.004	.005	.006	.007	.008	.009
Kilobars	.0	.1	.1	.1	.2	.2	.2	.3	.3

Note.—The value for gravity is that of the United States Coast and Geodetic Survey. A value 980.665 given by the Bureau of Standards was adopted in 1888 by the International Committee on Weights and Measures and has since been continued for convenience although it is a conventional standard and not exactly equal to the value at 45°. There has been a slight change in the value for the density of mercury. The differences are small. See *Monthly Weather Review* for April, 1914, p. 230, article by R. N. Covert.

Table 8. Millimeters to Kilobars

Millimeters	Kilobars	Millimeters	Kilobars
1	1.3	500	666.6
10	13.3	600	799.9
100	133.3	700	933.2
200	266.6	800	1066.6
300	400.0	900	1200.0
400	533.3	1000	1333.3

Millimeters to Kilobars

Base 1000 kbs. or one million dynes

Millimeters	0	2	4	6	8
700	−66.8	−64.1	−61.4	−58.8	−56.1
710	−53.4	−50.8	−48.1	−45.4	−42.8
720	−40.1	−37.4	−34.8	−32.1	−29.4
730	−26.8	−24.1	−21.4	−18.8	−16 1
740	−13.4	−10.8	− 8.1	− 5.4	− 2.8
750	− 0.1	+ 2.6	+ 5.2	+ 7.9	+10.6
760	+13.2	+15.9	+.8.6	+21.2	+23.9
770	+26.6	+29.2	+31.9	+34.6	+37.2
780	+39.9	+42.6	+45.2	+47.9	+50.6
790	+53.3	+55.9	+58.6	+61.2	+63.9

TABLE 9. CONVERSION OF TEMPERATURES

Absolute (A.°)	Centigrade (C.°)	Fahrenheit (F.°)	Réaumur (R.°)	Absolute (A.°)	Centigrade (C.°)	Fahrenheit (F.°)	Réaumur (R.°)
373	100	212	80	303	30	86	24
68	95	203	76	2	29	84	23
63	90	194	72	1	28	82	22
58	85	185	68	300	27	81	22
53	80	176	64				
48	75	167	60	300	27	81	22
43	70	158	56	299	26	79	21
38	65	149	52	98	25	77	20
33	60	140	48	97	24	75	19
28	55	131	44	96	23	73	18
325	52	126	42	295	22	72	18
24	51	124	41	94	21	70	17
23	50	122	40	93	20	68	16
22	49	120	39	92	19	66	15
21	48	118	38	91	18	64	14
320	47	117	38	290	17	63	14
19	46	115	37	89	16	61	13
18	45	113	36	88	15	59	12
17	44	111	35	87	14	57	11
16	43	109	34	86	13	55	10
315	42	108	34	285	12	54	10
14	41	106	33	84	11	52	9
13	40	104	32	83	10	50	8
12	39	102	31	82	9	48	7
11	38	100	30	81	8	46	6
310	37	99	30	280	7	45	6
9	36	97	29	79	6	43	5
8	35	95	28	78	5	41	4
7	34	93	27	77	4	39	3
6	33	91	26	76	3	37	2
305	32	90	26	275	2	36	2
4	31	88	25	74	1	34	1

TABLE 10. SUPPLEMENTARY TO TABLE 9
CONVERSION OF MINUS TEMPERATURES

Absolute (A.°)	Centigrade (C.°)	Fahrenheit (F.°)	Réaumur (R.°)	Absolute (A.°)	Centigrade (C.°)	Fahrenheit (F.°)	Réaumur (R.°)
273	0	32	0	240	−33	−27	−26
72	−1	30	−1	39	−34	−29	−27
71	−2	28	−2	38	−35	−31	−28
				37	−36	−33	−29
270	−3	27	−2	36	−37	−35	−30
69	−4	25	−3				
68	−5	23	−4	235	−38	−36	−30
67	−6	21	−5	34	−39	−38	−31
66	−7	19	−6	33	−40	−40	−32
				32	−41	−42	−33
265	−8	18	−6	31	−42	−44	−34
64	−9	16	−7				
63	−10	14	−8	30	−43	−45	−34
62	−11	12	−9	29	−44	−47	−35
61	−12	10	−10	28	−45	−49	−36
				27	−46	−51	−37
260	−13	9	−10	26	−47	−53	−38
59	−14	7	−11				
58	−15	5	−12	225	−48	−54	−39
57	−16	3	−13	24	−49	−56	−39
56	−17	1	−14	23	−50	−58	−40
				22	−51	−60	−41
255	−18	0	−14	21	−52	−62	−42
54	−19	−2	−15				
53	−20	−4	−16	220	−53	−64	−42
52	−21	−6	−17	18	−55	−67	−44
51	−22	−8	−18	16	−57	−71	−46
				14	−59	−74	−48
250	−23	−9	−18	12	−61	−78	−49
49	−24	−11	−19				
48	−25	−13	−20	10	−63	−82	−50
47	−26	−15	−21	8	−65	−85	−52
46	−27	−17	−22	6	−67	−89	−54
				4	−69	−92	−55
245	−28	−18	−22	2	−71	−96	−57
44	−29	−20	−23				
43	−30	−22	−24	200	−73	−100	−59
42	−31	−24	−25	198	−75	−103	−60
41	−32	−26	−26	193	−80	−112	−64

TABLE 10. SUPPLEMENTARY TO TABLE 9—*Continued*
CONVERSION OF MINUS TEMPERATURES

Absolute (A.°)	Centigrade (C.°)	Fahrenheit (F.°)	Réaumur (R.°)	Absolute (A.°)	Centigrade (C.°)	Fahrenheit (F.°)	Réaumur (R.°)
188	−85	−121	−68	80	−193	−315	−154
183	−90	−130	−72	70	−203	−333	−162
				60	−213	−351	−170
173	−100	−148	−80	50	−223	−369	−178
163	−110	−166	−88	40	−233	−387	−185
153	−120	−184	−96				
143	−130	−202	−104	30	−243	−395	−193
133	−140	−220	−112	20	−253	−413	−201
				10	−263	−431	−209
123	−150	−238	−120	5	−268	−440	−213
113	−160	−256	−128	4	−269	−452	−215
103	−170	−274	−136	3	−270	−454	−216
100	−173	−279	−138	2	−271	−456	−217
90	−183	−297	−146	1	−272	−458	−218
				0	−273.09	−459.4	−218.8

CERTAIN TEMPERATURES ON THE ABSOLUTE SCALE

0°A. Professor K. Onnes in the Cryogenic Laboratory at Leiden found that at the temperature of nearly absolute zero, electrical resistance in conductors disappears. At 2° the electrical resistance of a thread of mercury was negligible.
1°A. Onnes, in an effort to obtain solid helium, obtained this temperature.
2°A. Maximum density of liquefied helium.
4°A. Boiling point of helium.
10°A. Effective temperature of space. At an elevation of 80 kilometers (50 miles) the temperature ranges from 5° to 10°A.
20°A. Boiling point of hydrogen.
79°A. Boiling point of nitrogen.
86°A. Boiling point of argon.
79°A. }
91°A. } Liquid air, according to proportion of oxygen.
90°A. Oxygen boils.
163°A. Alcohol freezes.

170°A. Critical temperature of air above which it cannot be liquefied.
181°A. Lowest air temperature recorded by means of sounding balloons (at Batavia, Java, November 5, 1913, at an elevation of 17,000 meters). Above this the temperature rose.
194°A. Obtained by Rotch with sounding balloon, June 25, 1905, at an elevation of 14,800 meters.
195°A. Carbon dioxide boils. Plant life ceases.
213°A. Lowest temperature recorded by Scott in his expedition to the south pole.
233°A. Mercury freezes.
273°A. Water freezes.
277°A. Water at maximum density.
287°A. Earth's mean temperature.
351°A. Alcohol boils.
373°A. Water boils.
504°A. Tin melts.
600°A. Lead melts.
1234°A. Silver melts.
1283°A. Temperature of boiling lava.
1336°A. Gold melts.
1356°A. Copper melts.
1698°A. Invar melts.
1773°A. Cast iron melts.
1800°A. Iron melts.
3723°A. Temperature of the electric arc.
4073°A. Temperature of positive crater of arc.
6000°A. Sun's temperature.
7800°A. True arc high pressure.

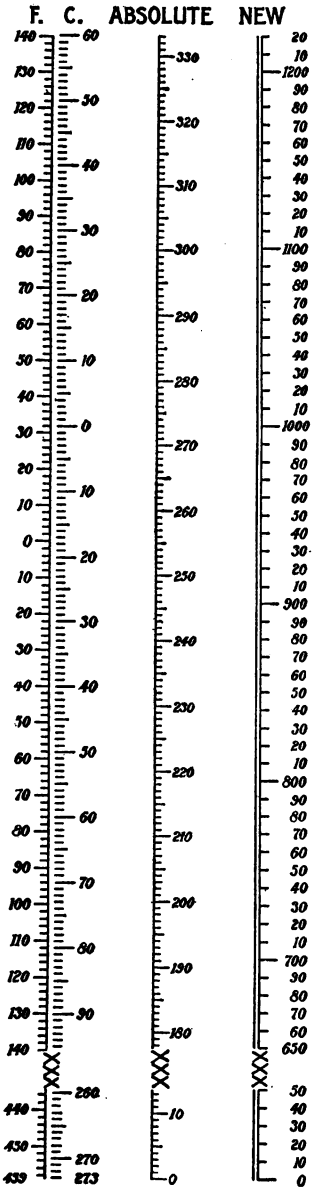


FIG. 114. TEMPERATURE SCALES

In Fig. 114 the three well-known temperature scales are charted side by side for quick comparison. There is also given a fourth scale, marked New, which is a variation of the Absolute Scale, the zero being the Absolute zero, and the 1000 degree mark the temperature of melting ice or 273°A. Like the Absolute Scale, this has no minus signs.

TABLE 11. BEAUFORT WIND SCALE

Beaufort number	Description	Pressure in kbs.	British Meteor. Office values (Miles/hour)	Weather Bureau values (Miles/hour)
0	Calm	.0	less than 1	0–3
1	Light air	.01	1–3	3–8
2	Light breeze	.04	4–7	8–13
3	Gentle breeze	.13	8–12	13–18
4	Moderate breeze	.32	13–18	18–23
5	Fresh breeze	.62	19–24	23–28
6	Strong breeze	1.1	25–31	28–34
7	Moderate gale	1.7	32–38	34–40
8	Fresh gale	2.6	39–46	40–48
9	Strong gale	3.7	47–54	48–56
10	Whole gale	5.0	55–63	56–65
11	Storm	6.7	64–75	65–75
12	Hurricane	8.	75–	75–

Beaufort wind scale

This is a scale adopted by Admiral Beaufort and used in connection with sailing vessels of the last century. It is unfortunate that it was ever seriously used by meteorologists. The most recent equivalent values are, as given by Galitzin, in accordance with the decisions of the International Meteorological Committee at the Rome meeting in 1913.

Force Beaufort scale	Meters per second	Force Beaufort scale	Meters per second
0	0	7	14–17
1	1	8	18–20
2	2–3	9	21–24
3	4–5	10	25–28
4	6–8	11	29–33
5	9–10	12	34 and over
6	11–13		

For the *c.g.s.* units the relations are, if f is the force upon a disk one square meter in area facing the wind, and v the velocity in meters per second,

$$f = 72\,v^2;$$

and if B is the Beaufort number, then

$$f = 4.78\,B^3$$

and $$v = 0.26\,\sqrt{B^3}.$$

TABLE 12. GRAMS OF AQUEOUS VAPOR PER KILOGRAM OF AIR AND VAPOR AT SATURATION

Temperature Absolute	Vapor pressure	Atmospheric pressure		
		1,000 kbs.	800 kbs.	400 kbs.
273	6.1 kbs.	3.5 grs.	4.7 grs.	9.5 grs.
274	6.6	3.8	5.1	10.2
275	7.1	4.1	5.5	11.0
276	7.6	4.4	5.8	11.8
277	8.1	4.7	6.3	12.7
278	8.6	5.0	6.7	13.6
279	9.3	5.4	7.2	....
280	10.0	5.8	7.7	....
281	10.7	6.2	8.3	....
282	11.5	6.7	8.9	....
283	12.2	7.1	9.5	....
284	13.1	7.6	10.2	....
285	13.9	8.1	10.9	....
286	14.9	8.7	11.6	....
287	15.9	9.2	12.4	....
288	17.0	9.9	13.2	....
289	18.0	10.6	14.1	....
290	19.0	11.2	15.0	....
291	20.3	12.0	16.0	....
292	21.6	12.8	17.0	....
293	23.2	13.6	18.0	....
294	24.6	14.5	19.4	....
295	26.0	15.0	21.0	....
296	28.0	16.0	22.0	....
297	29.0	17.0	23.0	....
298	32.0	18.0	25.0	....
299	33.0	20.0	....	....
300	35.0	21.0	....	....
301	37.0	22.0	....	....
302	40.0	23.0	....	....
303	42.0	25.0	....	....
304	44.0	26.0	....	....
305	47.0	28.0	....	....

Table 13. Grams of Aqueous Vapor per Cubic Meter at Saturation

°A.	grams	°A.	grams	°A.	grams
273	4.835	283	9.330	293	17.118
274	5.176	284	9.935	294	18.143
275	5.538	285	10.574	295	19.222
276	5.922	286	11.249	296	20.355
277	6.330	287	11.961	297	21.546
278	6.761	288	12.712	298	22.796
279	7.219	289	13.505	299	24.109
280	7.703	290	14.339	300	25.487
281	8.215	291	15.218	305	33.449
282	8.757	292	16.144	310	43.465

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